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1 **A litho-tectonic event stratigraphy from dynamic Late Devensian ice flow**
2 **of the North Sea Lobe, Tunstall, east Yorkshire, UK**

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14
15 **ABSTRACT**

16 The central sector of the British-Irish Ice Sheet during the last glaciation was characterised
17 by complex ice-flow reflecting interacting ice streams and changing dominance of different
18 ice dispersal centres. At Tunstall, east Yorkshire, two subglacial till units have been
19 traditionally identified as the Late Devensian Skipsea and Withernsea tills, and thought to
20 record two separate ice advances onto the Holderness coast, from divergent ice flow
21 directions. Our study presents the first quantitative lithological, sedimentological and
22 structural evaluation of glacial sediments at the site. The lithological composition of both till
23 units suggests that ice extended southwards from southern Scotland, incorporating material
24 from north-east England and the western margin of the North Sea Basin. Notably, the bulk
25 lithological properties of both the Skipsea and Withernsea tills are very similar. Subtle
26 variations in colour, texture and lithology that do occur simply appear to reflect spatial and
27 temporal variability in subglacial entrainment along the flow path of the North Sea Lobe.
28 The relative arrangements of the units plus the fracture sets also indicates phases of intra-till
29 thrust-stacking and unloading (F2), consolidation and shrinkage (F1, F3) suggestive of
30 cycles of ice re-advance (thrusting) and ice-marginal retreat (unloading and shrinkage)
31 possibly relating to active recession. The findings from this study reveal a sedimentary and
32 structural complexity that is not recognised by the current Late Devensian till stratigraphy of
33 east Yorkshire.

34

35 **KEYWORDS**

36 British–Irish Ice Sheet, North Sea Ice Lobe, Skipsea Till, Withernsea Till, Lithology,
37 Provenance

38

39 **1. INTRODUCTION**

40 Palaeo-ice sheets are important analogues for understanding contemporary ice sheets,
41 offering a record of ice sheet behaviour that can span millennia (Ely *et al.*, 2019). The last
42 British-Irish Ice Sheet (BIIS) provides an excellent analogue for understanding the character
43 and behaviour of modern marine-based ice sheets due to its comparatively small size,
44 accessible bed, and the wealth of pre-existing information (e.g. Evans *et al.*, 2005;
45 Livingstone *et al.*, 2012; McMillan and Merritt, 2012; Clark *et al.*, 2012; 2018).
46 Approximately two thirds of the BIIS was marine-based, drained by ice streams and fringed
47 by ice shelves in many places (Clark *et al.*, 2012; Gandy *et al.*, 2018, 2019), making it a
48 good analogue for the West Antarctic Ice Sheet (WAIS; Hubbard *et al.*, 2009). This is
49 especially so because the BIIS deglaciated in response to rising temperatures and a rising
50 sea level (driven by melting of other ice masses), which are the current forces that might
51 cause collapse of the WAIS (Bamber *et al.*, 2009; DeConto and Pollard, 2016).
52 Reconstructing the behaviour of palaeo-ice sheets enables a better understanding of the
53 long-term (centennial to millennial) behaviour of ice sheets within the Earth system. Only
54 when the mechanisms from palaeo-ice sheets, such as the BIIS, are better constrained, can
55 such knowledge be used for improving the next generation of numerical ice sheet models
56 used in sea-level forecasting (Stokes *et al.*, 2015; Ely *et al.*, 2019).
57 The body of empirical evidence related to the BIIS has progressively developed over the last
58 decade in particular (e.g. Clark *et al.*, 2012, 2018; Hughes *et al.*, 2016; Small *et al.*, 2017;

59 Bateman *et al.*, 2018; Bradwell *et al.*, 2019; Davies *et al.*, 2019; Ely *et al.*, 2019; Lovell *et*
60 *al.*, 2019), producing an ever-expanding database of palaeo-ice sheet data, but there are still
61 gaps in knowledge regarding ice-marginal processes (Roberts *et al.*, 2013) and their
62 implications for glacier dynamics. Eastern England and the North Sea Basin are the main
63 areas of complexity and uncertainty due to multiple competing ice lobes and potential ice
64 flow reversals (Evans *et al.*, 2019). The ice limits, interactions between ice lobes, and their
65 relative chronologies in this area are only broadly known and, without this understanding,
66 the glaciodynamic history and the nature of the BIIS remains contested.

67 In order to address outstanding questions on the character and behaviour of part of the
68 southeast sector of the last BIIS, this study has three main aims. Firstly, this study will
69 determine the depositional processes and a relative event stratigraphy for the Late
70 Devensian sediments observed at Tunstall, eastern England (**Figure 1**). Secondly, this study
71 will determine the provenance of the glaciogenic sequence in order to reconstruct the glacial
72 transport pathway for the deposits. Thirdly, this study will integrate this event stratigraphy
73 and sediment provenance information into the broader Late Devensian evolution of the
74 region and southeast sector of the last BIIS.

75

76 **2. STUDY AREA AND PREVIOUS INVESTIGATIONS**

77 A southward advance of the North Sea Lobe (NSL) of the last BIIS during the Late
78 Devensian (Weichselian; Dimlington Stadial; MIS 2) resulted in the development of till
79 sequences across parts of County Durham, east Yorkshire, Lincolnshire and north Norfolk
80 (Pawley *et al.*, 2006; Catt, 2007; Davies *et al.*, 2009, 2013; Boston *et al.*, 2010; Evans and
81 Thomson, 2010; Bateman *et al.*, 2011, 2015, 2017; Roberts *et al.*, 2013). The glacial
82 lithostratigraphy of the Holderness coast, east Yorkshire (**Figure 1B**), has traditionally been

83 subdivided into three till units (the Basement Till, Skipsea Till, and Withernsea Till
84 respectively), based principally upon particle size distribution, sediment colour, clast
85 lithology, heavy mineral composition and matrix calcium carbonate content (Madgett and
86 Catt, 1978; Catt, 2007). Optically Stimulated Luminescence (OSL) ages constrain the timing
87 of initial advance of the NSL into the region, and deposition of the Skipsea Till, to ~21.7 ka
88 (**Figure 1**), reaching its maximum extent at ~19.5 ka (Evans *et al.*, 2019) with offshore
89 retreat occurring at ~18 ka (Bateman *et al.*, 2011, 2015, 2017). The recession of the NSL
90 was initially rapid followed by a series of near synchronous oscillations of the NSL, and
91 subsequent deposition of the Withernsea Till at ~16.8 ka, before the final terminal retreat of
92 the ice sheet occurring prior to ~15.5 ka (Bateman *et al.*, 2017).

93 Catt (2007) described the Skipsea Till as possessing a dark-greyish brown (10YR 3/2)
94 colour and occupying the whole Holderness area. The Skipsea Till is correlative with the
95 Late Devensian Horden Till Formation in County Durham (Davies *et al.*, 2009; 2012), based
96 on stratigraphic relationships and sedimentary petrography. According to Pawley *et al.*
97 (2008), on the basis of correlative luminescence dates, the north Norfolk equivalent of the
98 Skipsea Till is the Holkham (Hunstanton) Till. However, recent OSL dating has suggested a
99 slightly earlier incursion of ice into north Norfolk timed at ~21.5 ka (Evans *et al.*, 2019).

100 Bisat (1939) and Radge (1939), suggested that the Skipsea Till was deposited by ice flowing
101 into Eastern England through the Stainmore Gap from the Lake District and Pennines, while
102 the Withernsea Till originated from the Cheviots and Southern Uplands. Later studies
103 documented a suite of erratics derived from Scotland, Northumberland, and the Cheviots
104 (Catt and Penny, 1966; Madgett and Catt, 1978; Catt, 2007). Busfield *et al.* (2015)
105 confirmed, from quantitative clast data and derived microfossils, that the Skipsea Till was
106 sourced from southern Scotland, incorporating material from north eastern England,
107 northeast Yorkshire and the western margin of the North Sea Basin.

108 The overlying Withernsea Till (dark-reddish brown, 5 YR 3/4 in its weathered form) by
109 contrast is less widespread, cropping-out in south east Holderness (**Figure 1C**) (Evans and
110 Thomson, 2010). North of Flamborough Head, the Withernsea Till reappears as the Upper
111 Till Series of Edwards (1981, 1987), seen in coastal cliffs of Filey Bay. However, Bisat
112 (1939) suggested that the till is confined to isolated basins within this area. The Withernsea
113 Till can be less confidently provenanced to its source area although erratics recorded
114 suggest a source from the Lake District and Pennines (Catt and Penny, 1966; Catt and
115 Digby, 1988; Bell and Forster, 1991). The red colouration is also strongly suggestive of
116 input of Permo-Triassic materials from the Sherwood Sandstone and/or Mercia Mudstone
117 groups. The fragmented, crenulated ridges developed within the landscape of Holderness
118 (**Figure 1C**; Evans *et al.*, 2001; Evans and Thomson, 2010; Clark *et al.*, 2018) and
119 superimposed upon the Skipsea and Withernsea Tills have been interpreted as lateral
120 moraines and also suggest an ice-flow direction from the north and east.

121 Using geochemical data from samples at seven sites along the Holderness coast, Boston *et*
122 *al.* (2010) argued that the Basement, Skipsea and Withernsea Tills could not be statistically
123 differentiated, with more variation within than between the tills. This raised significant
124 questions regarding the stratigraphical correlation of Late Devensian tills in east Yorkshire
125 and Lincolnshire. Boston *et al.* (2010) concluded that the till units are not lithologically
126 distinct and that the current stratigraphy does not recognise the sedimentary and structural
127 complexity produced by repeated onshore, possibly surging flow by a dynamic NSL along
128 the eastern margin of the BIIS (based on stacked sequences; Evans and Thomson, 2010).

129 Clast lithology has been used effectively to quantify ice-flow pathways and provenance for
130 the Skipsea Till (Busfield *et al.*, 2015), but no comparative work has yet been undertaken
131 for the Withernsea Till.

132 With some of the most rapidly eroding coastline in northern Europe (Bird, 2010; Castedo *et*
133 *al.*, 2015), the Holderness coast provides an ideal opportunity to assess extensive cliff
134 exposures through the Quaternary geology. Tunstall is centrally located within the limits of
135 both the Skipsea and Withernsea tills (**Figure 1C**), offering a valuable opportunity for a
136 comparative study between the two tills deposited in superposition. Excluding a
137 comprehensive soil profile examination by Madgett and Catt (1978) (**Figure 1D**), there has
138 been little previously published process or provenance work conducted at Tunstall. Due to
139 this lack of quantitative and detailed analysis, the understanding of till genesis remains
140 underdeveloped. The site at Tunstall, therefore, offers an excellent opportunity to use
141 detailed clast lithological analysis and detailed process-based sedimentology to untangle the
142 stratigraphic relationships, ice-flow pathways and provenance of the Skipsea and
143 Withernsea Tills.

144

145 **3. METHODS**

146 We present new quantitative sedimentological investigations of the Late Devensian till
147 sequence that crops-out above Cretaceous chalk bedrock at Tunstall, east Yorkshire (**Figure**
148 **1D**), together with a comprehensive clast lithological provenance analysis critical to the
149 palaeoglaciological reconstruction of the eastern sector of the BIIS.

150 Eight exposures were logged in detail from 2 km of vertical coastal cliff sections at Tunstall,
151 east Yorkshire (0°0'6.9"E, 53°45'24.7"N). At each site, vertical profile logs were compiled
152 from cleared sections. Following procedures outlined in Evans and Benn (2004), the
153 sedimentary characteristics were recorded including unit thickness, modal grain size,
154 sedimentary and tectonic structures, degree of consolidation, matrix *vs* clast supported
155 nature, grading and sorting of each unit, Munsell colour, bed geometry, and the nature of

156 contacts between the units. Sediments were described using standard facies codes (following
157 Benn and Evans, 1998) and reclassified into lithofacies associations (LFAs) to aid regional
158 correlation. In order to convey the lateral changes in architecture and localised complexity
159 in structural features, detailed field sketches and cross sections, supported by photographic
160 evidence, were utilised to accurately map the geometry of the exposures and create an
161 overall facies architecture map.

162 A three-order clast morphological analysis was used, encompassing clast shape,
163 angularity/roundness and stone orientation to help establish the depositional history of the
164 sediment (cf. Evans and Benn, 2004; Hubbard and Glasser, 2005; Hambrey and Glasser,
165 2012). Where possible, palaeo-current measurements were taken on stratified deposits. Clast
166 fabrics, striae measurements and eigenvector analyses followed Benn (1994) and Hubbard
167 and Glasser (2005). The structural data is presented in equal area stereographic projections
168 as poles to planes. Bulk samples for particle size analysis (PSA) and clast lithological
169 analysis (CLA) were collected from a 2 m² area in each lithofacies to give a statistically
170 significant, representative sample (Bridgland, 1986). The minimum sample size was 201
171 clasts; however, >300 clasts were counted in 4 of the 8 samples. At least two replicate
172 samples from each lithofacies for both PSA and CLA were taken to ensure the heterogeneity
173 within the stratigraphy was accounted for. Due to the spatial variability and clast-poor
174 nature of the diamicton units, bulk samples of at least 10 kg were collected. PSA was
175 undertaken in order to describe the textural properties of the sediments and support genetic
176 interpretation (Gale and Hoare, 2012). PSA was conducted using a laser granulometer for
177 particles <2000 µm and dry sieving for particles >2000 µm. The lithology of all recovered
178 clasts in the 8-16 mm, 16-32 mm, and >32 mm size fractions was identified using a low-
179 powered binocular microscope and compared to a reference collection and standard rock

180 identification criteria (Evans and Benn, 2004, Walden, 2004, Stow, 2005; Gale and Hoare,
181 2012).

182 Multivariate statistical analysis was performed on the clast lithological data using Principal
183 Component Analysis (PCA). PCA is commonly used in both regional geochemical and
184 lithostratigraphical studies (Gibbard, 1985; 1986; Cheshire, 1986; Scheib *et al.*, 2011). The
185 data are easily reduced into a smaller number of interrelated groups that reveal underlying
186 patterns or ‘principal components’ within lithological datasets. PCA axes simplify and
187 represent variation in the data (Davis, 1986) to identify key variables, and relationships
188 between variables, within a dataset (Richards, 1998; Lee, 2003; Davies *et al.*, 2009). The
189 analysis was run using the covariance matrix method (Kovach, 1995) which is strongly
190 affected by non-normally distributed data and outliers (Reimann *et al.*, 2008). Like other
191 exploratory multivariate statistical techniques, PCA provides eigenvectors, or ‘principal
192 component coefficients’ to describe the relative significance of individual (lithological)
193 components and their variability within the data set. Associated eigenvalues or ‘principal
194 component scores’ record the percentage of the total variance of each principal component,
195 in this case, measuring the importance of each lithology relative to each other in describing
196 the characteristics of a particular sample.

197

198 **4. RESULTS: SITE SEDIMENTOLOGY AND STRATIGRAPHY**

199 *4.1 Facies Architecture*

200 The stratigraphic architecture of the exposures at Tunstall is summarised in **Figure 2**.
201 Vertical profile logs 1 - 8 (**Figure 3a, b**) reveal four distinct Lithofacies Associations
202 (LFAs). Bedrock is not exposed and the LFAs identified vary both vertically and laterally.
203 The summarised relative stratigraphic succession observed in the field is as follows; dark

204 coloured diamict (LFA 1) exposed towards the north of the section, overlain by red-coloured
205 diamict (LFA 2), which is in-turn overlain by clast-supported sands and gravels with a
206 sharp, unconformable, undulating base (LFA 3). A distinctly white organic silt (LFA 4)
207 crops-out intermittently for a short distance within the LFA 3 succession in the middle of
208 the lateral section. Above LFA 4 is a thin unit of LFA 3b and the top of the sequence is
209 capped by a thin soil horizon ~10 cm thick.

210 The surface of the cliff undulates, and the middle section has been heavily human-modified.
211 The height of the section varies between 4 m at the lowest point, where there is a break in
212 the cliffs for beach access, to ~12 m at the highest. Slumping along the length of the section
213 is widespread due to the instability of the cliffs obscuring many of the *in-situ* sediments. An
214 unusual feature of the section at Tunstall is the numerous natural cliff promontories that
215 protrude out ~ 3m towards the sea. Coastal erosion is non-uniform and some spurs are more
216 prominent than others. An even distance of 10 - 15 m separates the spurs from each other
217 along the beach.

218 *4.2 Lithofacies Descriptions*

219 *LFA 1*

220 LFA 1 is a dark brown (7.5 YR 3/4), matrix-supported, massive and homogenised diamicton
221 with a dense clay-matrix texture, containing clasts ranging from fine gravels to small
222 boulders up to 25 cm in diameter (**Table 1**). It has an over-consolidated nature. The
223 thickness of LFA 1 varies between 0 - 6 m and is exposed only in the northernmost part of
224 the section (**Figure 2; Figure 3b**; Section logs 5 to 8). The majority of clasts within LFA 1
225 are sub-angular (34 %) to sub-rounded (38 %) (**Figure 6B**), incorporating a high proportion
226 of faceted and striated clasts (**Figure 6E**). The overall clast morphology is dominated by
227 'blocky' shapes (**Figure 6C**). The diamicton contains numerous laterally-extensive thin (<5

228 cm) lenses of gravel that dip 30 ° to the south and can be traced along the section in several
229 places (**Figure 3b**; Section logs 7 and 8). It possesses a generally weak clast fabric at each
230 locality sampled (**Figure 2**); with a polymodal distribution of points but clustering towards
231 the southeast. Stringers also occur in the lower few metres of the unit, just above the beach
232 level, persisting up to 1 m before pinching out. LFA 1 is fissile, observed particularly at the
233 boundary between LFA 1 and LFA 2. The upper contact with LFA 2 is sharp with a concave
234 base. The base of the LFA 1 was obscured by the beach.

235 *LFA 2*

236 LFA 2 is a massive, matrix-supported, bright reddish brown (5 YR 5/6) sandy diamicton
237 (**Table 1**). LFA 2 is consistently exposed to the south of the section where the cliffs are ~11
238 m high (**Figure 2**). The average clast morphology of LFA 2 is ‘blocky’ (**Figure 6C**) with a
239 high proportion of faceted (31%) and striated (6%) clasts (**Figure 6E**). Larger boulders and
240 cobbles are observed at variable heights and orientations and are heavily striated with often
241 more than one striae orientation (**Figure 2**). The clast fabric is weak, with a polymodal
242 distribution of points but a fairly weak clustering from the northwest to the southeast
243 (**Figure 2**). The striae measurements show a clearer directional indicator, towards the south-
244 west, than the clast fabrics (**Figure 2**). The majority of clasts are sub-angular to rounded
245 (**Figure 6B**). LFA 2 is largely massive at the macroscale, but stringers also occur as
246 numerous streaks of red and black diamicton up to 10 cm thick (**Figure 4b**). Discontinuous
247 lateral beds (~30 cm) of grey diamicton alternate between thicker units (>50 cm) of the
248 brown diamicton (**Figure 4d, e**) with sharp contacts above and below the interbeds. Thin,
249 dipping sand and gravel laminations towards the northeast are observed at variable heights
250 within LFA 2. In Section log 5 (**Figure 3b**), cross-bedded gravelly sands were observed in
251 LFA 2 and a gravel pod that pinches laterally. The gravel pod is composed of laminated fine
252 sands and fine gravels that coarsen-upwards.

253

LFA 3

254 LFA 3 is best illustrated in Section Logs 1, 2, 4, 7, and 8 (**Figure 3a, b**). LFA 3 is laterally
255 variable, observed intermittently every few meters along the length of the exposure, often
256 capping the sequence just below the soil horizon. Overall, there are three lithofacies present
257 within LFA 3; planar laminated dark clays and silts (LF 3a), planar, cross-bedded fine sands
258 (LF 3b), and cobbles and gravels (LF 3c). The basal contact of LFA 3 is undulating, with
259 elevations ranging from 1.2 m to 8 m. It is repeatedly down-cut into the underlying
260 diamicton (LFA 1 or 2) in a concave-shaped depression and has a sharp, erosive base (e.g.
261 **Figure 3a**; Section log 1). The upper contact with either LFA 4 or the soil horizon is planar
262 and sharp.

263 LF 3a is dominated by horizontal laminated dark clays and silts, which coarsen upwards into
264 massive fine sands (LF 3b). Thin (<5 cm) laterally-continuous seams of coal fragments and
265 particles (mm – cm) occur within the sandy unit of LF 3b. LF 3c consists of well-sorted,
266 well-rounded, clast-supported coarse pebbles and cobbles (**Table 1**) that are stratified and
267 imbricated towards the northwest (**Figure 3a**; Section Log 4; **Figure 4g, h**). Randomly
268 orientated, clast-supported, coarse gravels of LF 3c typically infill the concave structures
269 (**Figure 4g**) and numerous pebble lags are observed along the bases of the channel-like
270 structures. The majority of clasts are well-rounded (28 %), rounded (32 %) and sub-rounded
271 (24 %), with only 4 % faceted and no clasts striated (**Figure 6B, C, E**). Up to 14 % of the
272 stones are broken. The unit thickness of LF 3c is typically thicker than LF 3a and b.

273

LFA 4

274 LFA 4 is laterally discontinuous and only crops-out within LFA 3 near the surface of the
275 southernmost part of the section (**Figure 4f**). LFA 4 comprises a bright white (7.5YR 8/1)
276 silt which highly calcareous (57 % calcium carbonate content). The contacts above and

277 below LFA 4 are very sharp and linear. The unit of the lower boundary has a slight elliptical
278 shape covering a discontinuous lateral extent of roughly 300 m. The deposit is uniformly 20
279 cm thick and is densely packed with small shells deposited in their life position. The
280 majority of the shells within the sediment are freshwater gastropods including abundant
281 *Radix bulthica*, *Lymnaea peregra*, and *Galba truncatula*, but other freshwater snails
282 *Planorbis viviparis*, *Valvata crisata* and aquatic bivalves of *Pisidium* genus are also present.

283 *4.3 Lithofacies interpretations*

284 *LFA 1 and 2*

285 LFA 1 and LFA 2 possess many bulk characteristics of a subglacial traction till such as their
286 matrix-supported texture, highly fissile, and over-consolidated nature, in addition to the
287 presence of faceted and striated lithologies of wide-ranging provenance (cf. Evans *et al.*,
288 2006). A subglacial traction till is defined by Evans *et al.* (2006) to include sediments that
289 accreted by sliding over and/or deforming at the glacier bed, sediment released directly from
290 the ice by pressure melting, and sediment completely or largely homogenised by shearing.
291 The fissile structures, observed at the boundary between the tills are interpreted here to be
292 very small, thin thrust faults. Clast rotation within subglacial traction tills results in shape
293 alignment of elongate, low-sphericity grains such as those observed in LFA 1 and 2 (**Figure**
294 **6B, C**). Where clasts have been brought into contact while lodged on a rigid bed and
295 moving in the subglacial traction zone, they typically have bullet-shaped and faceted ends
296 (Boulton and Hindmarsh, 1987; Kruger, 1979, 1984, 1994; Sharp, 1982), such as those in
297 LFA 1 and 2 (**Figure 6E**). Other structures within both lithofacies indicative of lodgement,
298 such as their massive homogenised appearance, occur in association with structures
299 indicative of subglacial deformation, including stringer initiation and deformed inclusions of
300 LFA 1 in LFA 2 (**Figure 4b, d, e**), as well as weak clast macrofabrics (**Figure 2; Figure 3a,**
301 **b**); Evans *et al.*, 1995; Hicock and Fuller, 1995, Hart 1997; Bennett *et al.*, 1999; Roberts and

302 Hart, 2005). Stringers of red and black diamicton (particularly in LFA 2) are, therefore,
303 interpreted as tectonic laminae produced by the progressive shearing and attenuation of soft
304 sediment inclusions (Hart and Roberts, 1994; Phillips *et al.*, 2008). Most of these structures
305 indicate that the till was formed under low strain conditions with elevated porewater
306 pressures (van der Wateren, 1995; Hiemstra *et al.*, 2007; Lee and Phillips, 2008); however,
307 strain rates can vary both spatially and temporally as pore-water pressure fluctuates, creating
308 a mosaic of deformation (Piotrowski *et al.*, 2004; Lee and Phillips, 2008; Lee, 2009). This
309 may explain why some of the sections at Tunstall, particularly of LFA 1, are more massive
310 and are completely homogenised than others, while delicate deformation structures, such as
311 stringers, are preserved elsewhere.

312 *LFA 3*

313 Overall, a glaciofluvial origin is suggested for LFA 3 based upon the coarse-grained
314 characteristics of LF 3c in addition to the evidence for abrupt discharge fluctuations,
315 recorded in discontinuous, lensate bodies of cross-bedded sands (LF 3b) which are likely to
316 be post-glacial winnowed lags. These packages of massive to crudely horizontally-bedded
317 sheets, separated by lower discharge scour infills, are typical of strongly episodic fluvial
318 sedimentation, classified as gravel sheets by Miall (1977) and Maizels (1993). The
319 horizontally-bedded glaciofluvial outwash assemblage records rapidly fluctuating
320 discharges, as evidenced by abrupt vertical changes from boulder gravels to the laminated
321 sediments typical of overbank fines or waning discharge drapes (Miall, 1977, 1985;
322 Collinson, 1996).

323 The irregular, concave base of the LFA 3 basal contact is interpreted as an erosional base
324 with concave-up bases and flat tops (**Figure 4g**) permitting isolated channelized forms that
325 have been shaped by the erosive force of water incising into the underlying sediments. The
326 gravel facies that infill the isolated channel forms, in addition to the strong variation in

327 height of these channels, support the interpretation of deposition in a proximal setting for
328 LFA 3, interpreted as the product of high energy proglacial outwash when the ice was
329 retreating (Miall, 1977; Maizels, 1995). Due to the lack of large-scale trough cross-bedding
330 and the cyclic fining-upwards sequence of gravels, sands and silts, distal proglacial outwash
331 sedimentation has been excluded as a mechanism for the deposition of LFA 3. Instead, LF
332 3a could have been deposited by under-melt at the ice-bed interface in subglacial canals (cf.
333 Walder and Fowler, 1994), perhaps indicative of ice-bed decoupling and sliding, or in Nye
334 channels, similar to those that have been observed elsewhere in Devensian sediments in
335 Northeast England (Eyles *et al.*, 1982; Davies *et al.*, 2009). The pebble lags evident in LFA
336 3a point to evidence of bedload saltation, whereby pebbles have been rolled along in
337 flowing water at the boundary of the bed and formed a lag.

338 The direction of the imbricated clasts of LF 3c indicates a palaeo-flow direction from the
339 northeast. The majority of clasts in LFA 3c are rounded (**Figure 4b**; **Figure 6B**) which
340 indicates highly abrasive high energy conditions in a fast-flowing current, whereas the
341 lenses of bedded sands, and stratified gravels are indicative of moderate flow regimes with
342 frequent changes in flow regime and sediment supply. An increasing energy regime is also
343 implied by the upwards coarsening of the sequence. Changes in the dominance of flow
344 suggest the presence of fast, hyperconcentrated flows, as well as slower moving, lower
345 energy flows.

346 *LFA 4*

347 The freshwater gastropods *Radix bulthica*, *Radix peregra*, *Galba truncatula*, *Panorbis*
348 *viviparous* and *Valvata cristata* all inhabit the same type of environment; stagnant or slow
349 moving water (Kerney *et al.*, 1980). Due to the abundance of these freshwater gastropods,
350 the depositional environment is inferred to be a spring-fed vegetated calcareous pool or
351 shallow film of water trickling across wet ground such as a Tufa (Garnett *et al.*, 2004). Tufa

352 is a variety of limestone formed by the precipitation of carbonate minerals from ambient
353 temperature water bodies and forms either in fluvial channels or lacustrine settings. The
354 deposit at Tunstall is likely to have occurred from an emergence of a spring or seep, due to
355 its thin bed thickness and discontinuous lateral extent. Spring-fed paludal deposits are
356 widespread on or near limestone bedrock (Andrews *et al.*, 2000).

357 *4.4 Structural genesis*

358 *Fracture description*

359 A series of horizontal and vertical fractures dissect the lower sections of the cliffs at
360 Tunstall. In places, these fractures have been eroded and enlarged by wave action. Overall,
361 four distinct fracture sets are observed within the cliff exposures; three sub-vertical sets (F1,
362 F3 and F4), punctuated by a sub-horizontal fracture set (F2). Given their continuous
363 presence and abundance along the cliff exposure, the fractures are inherent structural
364 features of the deposits at Tunstall.

365 The F1 fracture set occurs discontinuously at the base of the cliffs within LFA 1 material.
366 They are vertical to sub-vertical fractures, up to 1.5 m in length, that strike broadly
367 northeast-southwest and variably dip towards the northwest and southeast at angles $>75^\circ$
368 (**Figure 5**). The density of the fractures is spatially variable, being more densely
369 concentrated at the base of each cliff promontory but occurring along the entire length of the
370 described section (**Figure 4a, c, d, i**). The density of fractures varies with zones of high
371 density (spacing ranging from 1 - 20 cm) and low density (spacing range from 20 - 100 cm)
372 fracturing. The F1 fracture set extends up to, and is truncated by, a large persistent
373 horizontal fracture (F2).

374 Horizontal fracture, F2, forms a laterally-persistent horizon, truncating F1 and separating
375 beds of lower and upper unit of LFA 1 (**Figure 4c, i**). F2 can be traced discontinuously

376 along the cliff sections typically up to 2 m above the level of the foreshore. Dip and dip
377 azimuth measurements collected on the fracture surfaces demonstrate a spread polymodal
378 distribution with shallow ($<14^\circ$) dips (**Figure 5**) indicating a gently undulating sub-
379 horizontal fracture plane.

380 The F3 fracture set extend upwards from the sub-horizontal fracture F2 occurring within the
381 upper unit of LFA 1 and also in LFA 2, but are cross-cut in-turn by LFA 3 and 4. This
382 demonstrates, that F3 post-date F2 but pre-date the deposition of LFA 3 and 4. They are
383 visible discontinuously along the length of the section, and are sub-vertically inclined ($>78^\circ$)
384 and have a marked straightness and parallelism (**Figure 5**). They range in length from 10 cm
385 to 4 m and strike broadly east northeast - west southwest.

386 Fractures F4 occur only within the cliff promontories where they truncate the entire vertical
387 section (including F1-F3) with their bases occurring beneath beach level. The fractures are
388 slightly curved fractures that radiate upwards, sub-parallel to the margins of the
389 promontories. The fractures widen upwards (up to 2 m) and towards the promontory
390 margins.

391 *Fracture Interpretation*

392 Fractures F1-F3 are interpreted as extensional fractures (mode 1 fractures) due to the lack of
393 evidence for slip or displacement along the joints. Collectively they are interpreted to form
394 broadly perpendicular sets of sub-vertical (F1 and F3) and horizontal (F2) joints formed by
395 glacier unloading, consolidation and drying of the diamicton. Sub-horizontal joints (F2) are
396 unloading joints (also called release joints) formed as the applied vertical load (ice
397 overburden) was removed. The removal of the overburden caused the vertical compressive
398 stress to be released resulting in fracturing being initiated probably along a pre-existing
399 plane of weakness. The continuous nature of F2 implies that this pre-existing plane of

400 weakness was laterally extensive, for example a shear plane, which bounded two beds of
401 LFA 1.

402 The origin of vertically-aligned fractures (F1 and F3) is also interpreted to be an artefact of
403 unloading but also subsequent drainage, drying (Boulton and Paul, 1976) and shrinkage of
404 the sediment (Mertens *et al.*, 2003). Within this scenario, removal of the ice overburden led
405 to a reduction in the differential stress and a switch from dominant vertical to horizontal
406 compression (Maltman, 1994). Subsequent reduction in the horizontal compression and a
407 switch to tensile stresses would have promoted shrinkage which in-turn would require a
408 moderate differential stress to be maintained to prevent shearing (Mertens *et al.*, 2003;
409 Dehandschutter *et al.*, 2005).

410 F4 fractures are focussed around the coastal spurs or promontories and cross-cut all other
411 parts of the sequence demonstrating that these are the last features to have formed. Their
412 geometry, sub-parallel alignment to the margins of the promontories and the upwards-
413 widening of the fractures, suggests that they formed by the lateral release of an applied load
414 (i.e. the removal of cliff material by coastal erosion; Cossart *et al.*, 2008; Genter *et al.*,
415 2004).

416

417 **5. RESULTS: CLAST PROVENANCE**

418 *5.1 Clast lithological analysis*

419 Clast lithological data (**Table 2**) shows average percentages for each LFA. Eight samples
420 were analysed in total, four from LFA 1, three from LFA 2, and one sample from LFA 3.

421 **Figure 7** shows the results of clast lithological analysis for each sample. Representative
422 photographs are shown in **Figure 8**. The dominant lithologies within LFA 1 are

423 Carboniferous Sandstone (13.6 %), Jurassic Mudstone (11.7 %), Cretaceous chalk (10.4 %),

424 Carboniferous Limestone (9.5 %), Magnesian Limestone (6.6 %), and Jurassic Sandstone (6.
425 2 %), (**Figure 7**). Other clast lithologies within LFA 1 include Mercia Mudstone (3.02 %),
426 Old Red Sandstone (2.9 %), Sherwood Sandstone (2.7 %), and coal (2.1 %). There are
427 relatively low amounts of Whin Sill dolerite (1.9 %), greywacke (1.2 %), diorite (0.8 %),
428 Jurassic limestone (0.6 %), andesite (0.6 %), and rhyolite (0.5 %). Notably, there is a lack of
429 distinctive Lake District granites and erratics.

430 LFA 2 contains higher percentages of Carboniferous sandstone (18.4 %), undistinguished
431 arkosic sandstone (4.6 %), Whin Sill Dolerite (4.3 %) and Sherwood Sandstone (3.1 %),
432 than LFA 1, but lower amounts of Carboniferous limestone (7.4 %), Jurassic mudstone (7.0
433 %) and Cretaceous chalk (6.6 %). There are slightly higher percentages of igneous
434 lithologies present within LFA 2, such as micro-granite (1.1 %), rhyolite (1.1 %), andesite
435 (1.0 %), basaltic porphyry (1.0 %), rhyolitic porphyry (0.6 %), andesitic porphyry (0.8 %)
436 and gabbro (0.3 %).

437 LFA 3 is dominated by a high proportion of Carboniferous sandstone (26.5 %) (**Figure 7**).
438 Other lithologies present in significant amounts are Jurassic mudstone (6.5 %), Old Red
439 Sandstone (6.2 %), yellow sandstone (5.4 %), red siltstone (5.0 %), and Jurassic sandstone
440 (3.1 %), plus an elevated amount of Whin Sill dolerite (9.6 %), much higher than is present
441 in either LFA 1 (1.9 %) or LFA 2 (4.3 %).

442 *5.2 Clast durability*

443 The two till units at Tunstall contain striated, faceted, far-travelled, non-durable erratics as
444 well as a matrix-supported sediment, indicating that a diverse suite of bedrock (including
445 softer lithologies of chalk and coal) have been eroded subglacially, cannibalised and
446 incorporated into the sediment. This suggests that the genesis of both tills was largely the
447 result of comminution and soft-sediment mixing (Roberts *et al.*, 2013). The mixed

448 lithological assemblages within LFA 1 and 2 suggest subglacial cannibalism from a number
449 of different regions. CLA confirms that both tills have not been deposited close to their
450 sediment provenance areas since their lithological composition is heterogenous, indicating
451 an increasing number of sources rocks over which the glacier has travelled (cf. Boulton,
452 1996a, b). It is unlikely that the abundance of soft, non-durable, sedimentary lithologies
453 (**Figure 9**) including Jurassic, Permian and Carboniferous limestone, Jurassic mudstone,
454 coal and chalk would have survived multiple episodes of re-working, particularly within a
455 highly abrasive, subglacial environment (cf. Lee *et al.*, 2002).

456 LFA 3 contains a diverse suite of erratics similar to those in LFA 1 and LFA 2 but there is a
457 noticeable paucity of non-durable lithologies such as Cretaceous chalk, Permian Magnesian
458 Limestone, and Jurassic Mudstone (**Figure 9**). The major lithological difference between
459 LFA 3 and LFA 1 and 2 is the marked increase in Whin Sill dolerite (~10 %) in LFA 3,
460 compared with <5 % in both LFA 1 and LFA 2. Almost the entire lithological content of
461 LFA 3 consists of far-travelled, durable lithologies (**Figure 9**). The presence of durable
462 lithologies support the interpretation of a high-energy environment as softer, non-durable
463 sedimentary lithologies are likely to have been destroyed by abrasion and attrition. The coal
464 fragments observed in LF 3b are an anomaly since coal is a low-durability material.
465 However, the relative buoyancy of coal aids its preservation during high-energy transport
466 whilst its deposition implies a rapid reduction (still-water) in energy regime (cf. Lee *et al.*,
467 2015).

468 *5.3 Principal Component Analysis (PCA)*

469 PCA was undertaken on the clast lithologies in an attempt to identify lithological similarities
470 between samples and determine the significant lithological variables and their
471 interrelationship. In total eight different principal components (PC 1 – 8) collectively
472 account for the lithological variability within the dataset. Three principal components (PC 1

473 – 3) account for 99.2 % of the total variability within the dataset. PC 1 explains 94.3 % of
474 the sample variation, whilst PC 2 accounts for 3.6 %, and 1.3 % is accounted for by PC 3.
475 **Figure 10a, b, c** shows the differences between lithofacies samples in relation to the
476 principal component scores (PC 1 – 3). Samples with similar clast compositions are
477 expected to cluster close together. PC 1 does not discriminate between the till samples (LFA
478 1 and 2) and the clast lithofacies (LF 3c) (**Figure 10a**). The samples do not cluster into
479 distinct groups, and the variation between samples in each LFA is large, indicating a broad
480 clast lithological spread throughout the samples. LFA 3 plots individually (**Figure 10a, c**),
481 due to the fact it has fewer non-durable lithologies (**Figure 9**). All other samples show
482 stronger variations along the axis of PC 1.

483 Based upon PC 1 there appear to be no major discernible differences in till lithology evident
484 between LFA 1 and LFA 2. However, there are subtle variations (inverse relationships)
485 between the other principal components (PC 2 and PC 3; **Figure 10b** and **c**). The principal
486 component scores suggest the correlation of samples with the established regional
487 lithostratigraphy (Skipsea and Withernsea Till) cannot be defined by bulk lithology (e.g. PC
488 1) mirroring results from other geochemical data analyses (Boston *et al.*, 2010). The
489 inability of the method to discriminate between the two tills could be a consequence of the
490 small sample number taken from each of the lithofacies as this can create an artificially high
491 skew. Additionally, the stratigraphic position of LFA 2 cropping-out above LFA 1 is also
492 likely to affect its lithological composition, since the bedrock would have already been
493 mantled with till, and a re-advance would at least locally cannibalise the underlying till. The
494 lithology of LFA 2 differs slightly, as shown by the more durable clast content (**Figure 9**).
495 PC 2 however, is able to discriminate between LFA 3 and the other till samples.

496 The principal component coefficients also identify key relationships between the main
497 source provinces for the clast lithologies (**Figure 10d, e, f**). PC 1 highlights the abundance

498 of Carboniferous and Jurassic material relative to other clast lithologies, accounting for 94.3
499 % of the lithological variability within the samples. PC 2 (3.6 %) shows an inverse
500 relationship between Carboniferous, Old Red Sandstone (positive), and Jurassic, Permian,
501 and Cretaceous lithologies (strongly negative). PC 3 (1.3 %) demonstrates an inverse
502 relationship between Carboniferous and Triassic lithologies (strongly positive) and Whin
503 Sill dolerite and Old Red Sandstone (strongly negative). Collectively, the principal
504 component coefficients demonstrate complex shifts in clast lithological variability
505 throughout the Tunstall sequence. Different processes relate to the suite of clasts present in
506 each LFA, particularly between the tills (LFA 1 and 2) and glaciofluvially transported
507 material (LFA 3). Whilst clast preservation during transportation is likely to play a
508 significant role, it is likely that the relationships observed reflect considerable temporal and
509 spatial variability (including intra-till) in the entrainment of materials from the subglacial
510 bed along the ice flow path. This may reflect a temporal and spatial partitioning within the
511 subglacial bed, both in terms of areas of bedrock cover and zones of subglacial erosion, with
512 areas of the subglacial bed either buried by younger superficial deposits or strain rates being
513 too low to drive erosion. A zonal approach to glacial erosion likely relates to the suite of
514 clasts found within the tills, whereas the lack of local lithologies within LFA 3 is likely a
515 result of high energy transport.

516 *5.4 Till provenance*

517 In both till units, LFA 1 and LFA 2, Carboniferous and Jurassic components constitute >50
518 % of the lithological composition of the samples (**Figure 7**). There has been an abundant
519 incorporation of Carboniferous sandstone, limestone and coal, which are characteristic of
520 the bedrock strata from North Yorkshire, County Durham and Northumberland (Taylor and
521 Eastwood, 1971; Jones *et al.*, 1995). Far-travelled sedimentary lithologies include the
522 Magnesian limestone (Cadeby Formation), which crops out extensively across north

523 Yorkshire, southern County Durham and offshore (Smith, 1995). Jurassic components are
524 traced back to outcrops in the regional area of the Cleveland basin and Redcar mudstone
525 formations in the Tees Bay (Macklin, 1998) and around Middlesbrough (Kent, 1980). Albeit
526 in low abundance in the tills, oolitic limestone is also typical of the Jurassic strata in the
527 northern part of the Cleveland basin. The Permo-Triassic Sherwood Sandstone Group
528 outcrops from the Tees Estuary to the Vale of York and the Midland Valley of Scotland
529 (Cameron and Stephenson, 1985).

530 Rare, far-travelled erratics in LFA 1 and LFA 2 include Lower Palaeozoic greywacke
531 (**Figure 8a**), which is likely to have been derived from the Southern Uplands (Greig and
532 Pringle, 1971). Devonian Old Red Sandstone clasts (**Figure 8c**) are likely to have a
533 provenance from the Midland Valley of Scotland (Trewin, 2002). Crystalline, metamorphic
534 lithologies such as schist, gneiss, diorite and k-feldspar rich granites and granodiorites
535 characteristic of the Scottish Grampian Highlands are present (**Figure 8g, j, k**), but in
536 extremely low abundance. This might suggest that these clasts have been re-worked from
537 the widespread Old Red Sandstone conglomerates that contain Dalradian material in the
538 Midland Valley (Cameron and Stephenson, 1985) or that their proportions have been diluted
539 by the incorporation of more local materials. K-feldspar granite clasts could also be derived
540 from the Cheviot Hills. Both tills contain purple to reddish-brown rhyolites, greenish-brown
541 andesites and porphyries (**Figure 8d, h**), which are characteristic of the Devonian Cheviot
542 Volcanic Formation that straddles the border between Northumberland and Scotland
543 (Toghill, 2011; Robson 1976). Rhyolite and andesite can also be attributed to the Lake
544 District, but the reddish-pink nature of the majority of rhyolites from Tunstall (e.g. **Figure**
545 **8h**) indicate a stronger presence of feldspar, characteristic of outcrops in the Cheviot Hills
546 (Robson, 1976). The Cheviot Hills is, therefore, established as the principal source region

547 for many of the igneous lithologies due to the large extent of felsic intrusive material in this
548 area.

549 The overall lithological composition of both tills (LFA 1 and LFA 2) suggest that they were
550 deposited by ice that flowed southwards down the present east coast of England. Source
551 regions for the ice, indicated by the clast lithological composition, include the Midland
552 Valley of Scotland, Southern Uplands and East Grampian Highlands, before ice entrained
553 erratics from the Cheviots, Northumberland and Durham (**Figure 11**) moving southwards to
554 the Holderness area.

555

556 **6. DISCUSSION**

557 *6.1 Implications for the Devensian Stratigraphy of eastern England*

558 LFA 1 and 2 are both massive, matrix-supported diamictons. At the macroscale, both units
559 appear to possess a similar texture, particle size distribution and clast content. A chi-square
560 test on the particle size distribution (**Table 1**) demonstrate that both tills cannot be
561 differentiated on the basis of particle size ($\chi^2_{\text{calc}}(0.002) < \chi^2_{\text{crit}}(0.35)$), meaning that they
562 could originate from the same source material. Other similarities between LFA 1 and 2
563 include laminations of gravel that are observed sporadically within both units and stringers
564 that also occur at the base of both LFA 1 and 2. However, there are also subtle meso-scale
565 (cm to m) variations between LFA 1 and 2 and distinguishing characteristics are matrix
566 colour and fracture density. Along the length of the section there is only one short exposure
567 where the lower two LFAs can be observed in superposition (**Figure 3b**; section log 5;
568 **Figure 4i**). At this boundary, the contact between the units is sharp and highly fissile. There
569 are also sand and gravel interbeds at the contact. The colour contrast at this boundary
570 between LFA 1 and 2 is sharp. LFA 1 is darker in colour, particularly at the base, due to

571 being constantly saturated with water from the sea at high tide. Fissile structures are more
572 abundant within LFA 1 than LFA 2, particularly at the contact where the upper contact with
573 LFA 2 is sharp with a concave base. LFA 1 is slightly more stone-rich and more heavily
574 fractured at the base than LFA 2 (**Figure 4i**).

575 Previous interpretations of the stratigraphic sequences along the Holderness coast have
576 suggested that two Late Devensian tills are present - the Skipsea and Withernsea tills (Catt
577 and Penny, 1966; Madgett and Catt, 1978; Bell and Forster, 1991; Bowen, 1999; Bell, 2002;
578 Catt, 2007). These classifications are founded largely on the basis of changes in matrix
579 colour and clast lithological assemblage (Catt and Penny, 1966; Madgett and Catt, 1978;
580 Bowen, 1999; Catt, 2007). In this study, LFA 1 is interpreted as a subglacial traction till
581 equivalent to the Skipsea Till on the basis of its dark brown colour, greater chalk content,
582 and northern clast erratic assemblage. This interpretation is regionally supported by other
583 investigations of tills which correlate stratigraphically to the Skipsea Till, such as the
584 Horden Till Formation at Whitburn Bay (Davies *et al.*, 2009), other areas of east Yorkshire
585 (Catt, 2007; Boston *et al.*, 2010; Evans and Thomson, 2010), Northumbria (Eyles *et al.*,
586 1982), and offshore (Carr *et al.*, 2006; Davies *et al.*, 2011). Based on its stratigraphic
587 position (overlying the Skipsea Till), LFA 2 should therefore be assigned to the Withernsea
588 Till. Sedimentologically, LFA 2 is similar to previous descriptions of the Withernsea Till; it
589 is a dark reddish brown sandy diamicton. However, the lithological data presented in this
590 study does not support the concept of LFA 2 being an entirely different till, as its bulk
591 lithology is indistinctive from the Skipsea Till.

592 The till sequences that crop-out along the east coast of England have consistently been
593 referred to as the product of either Scottish or east coast ice from the North Sea ice lobe
594 (NSL), or Stainmore ice. Clast lithological data from this study concludes that the tills
595 possess similar lithological characteristics and provenance. Evidence of Lake District input

596 (cf. Bisat, 1939; Radge, 1939; Catt and Penny, 1966; Catt and Digby, 1988; Bell and
597 Forster, 1991) was not replicated in this study and the lithological analysis at Tunstall
598 reveals no indicator erratics from the western part of England, now widely discredited
599 nevertheless (Davies *et al.*, 2019). This confirms that the tills are not associated with the
600 same ice lobe as the Vale of York glacier that formed the York and Escrick moraines
601 (Phillips, 1827; Howarth, 1903; Melmore, 1935, p. 31) as previously thought (Ford *et al.*,
602 2008).

603 The visual difference between the two tills at Tunstall can potentially be explained by the
604 local incorporation of rafts of the Sherwood Sandstone and /or Mercia Mudstone group (or
605 rafts of till units rich in these materials). Intermixing of the two tills (e.g. **Figure 4d, e**)
606 demonstrates that colour changes likely reflect subtle differences in till composition and
607 sediment source rather than weathering. Similar red diamictos and sands interbedded with
608 grey diamictos also occur within Devensian tills at Warren House Gill in Country Durham
609 (Davies *et al.*, 2012), indicating that this is a regional phenomenon. Whilst locally, this may
610 enable the apparent sub-division of till units into Skipsea and Withernsea Till facies, at other
611 sites to the north in Holderness (e.g. between Mappleton and Skipsea), facies of ‘Withernsea
612 Till’ occur within the ‘Skipsea Till’ (Jonathan Lee, unpublished data). Principal
613 Components Analysis also reveals subtle clast lithological variations within the till units
614 demonstrating greater intra- rather than inter- till variability. Therefore, on the basis of clast
615 lithological composition alone, it is difficult to discriminate between the two subglacial tills
616 identified at Tunstall. In simple terms, there is not a consistent superpositional relationship
617 between ‘Skipsea Till’ and ‘Withernsea Till’ facies in the Tunstall area and thus the
618 lithostratigraphic scheme becomes unviable. This lithological analysis supports the findings
619 from Boston *et al.* (2010), where geochemical analysis of LGM tills and glacioteconites in
620 east Yorkshire and Lincolnshire failed to precisely differentiate the Skipsea and Withernsea

621 Till types. This makes the application of the traditional nomenclature over a wider regional
622 area tenuous.

623 Given that the stratigraphic succession at Tunstall cannot be assigned to the traditional
624 bipartite sequence, we propose that inter- and intra-till variability relates to changes in
625 subglacial debris provenance. Variations in debris provenance can be explained by both
626 temporal and spatial changes in the availability of source materials implying that geological
627 sources cycled through phases of active and non-entrainment. This entrainment could occur
628 by melt and refreeze at the margins of the glacier prior to advance, by active shearing of
629 overridden stony permafrost, from supraglacial sources, or by normal melt-freeze
630 entrainment processes in temperate ice (Alley *et al.*, 1997) whereby the glacier actively
631 entrains basal material derived from the substrate by abrasion. To readily entrain debris into
632 the basal layers of glacial ice, debris entrainment encompasses the detachment of frozen
633 blocks of sediment from the subglacial substrate which is then folded and thrust. The
634 general entrainment mechanisms for basal debris transportation make these units
635 rheologically distinct. It is suggested that this behaviour may relate to changes in subglacial
636 conditions and behaviour. For example, the temporary burial (or exposure) of a specific
637 source material and/or changes in the subglacial bed rheology which drive stick (erosion)
638 and slip (non- or reduced entrainment) ice flow (Iverson, 2010; Iverson and Peterson, 2011;
639 Phillips *et al.*, 2018).

640 *6.2 Genetic Model: the significance of fractures F1 – F3*

641 Fractures F1-F3 are interpreted as being formed in response to unloading and shrinkage of
642 the tills in response to the removal of overlying glacier ice, followed by consolidation and
643 drying. F2 fractures are interpreted as unloading joints aligned perpendicular to the direction
644 of unloading (vertical). The geometry of the F2 fracture implies that the fracture developed
645 on a pre-existing, regionally-extensive plane of weakness such as a décollement surface.

646 The simplest interpretation is that this décollement surface originally formed due to the low-
647 angle glaciotectonic thrust emplacement of a layer of LFA 1 on top of LFA 1 (cf. Hiemstra
648 *et al.*, 2007; Lee *et al.*, 2013, 2017). The low-angle geometry of the décollement surface
649 implies that porewater pressures along the detachment were elevated (Phillips *et al.*, 2008;
650 Lee *et al.*, 2013, 2017). However, the sharpness of the fracture and absence of dewatering
651 structures (e.g. diffuse bedding, flame structures) suggests that a degree of consolidation and
652 dewatering of the lower unit of LFA 1 had occurred prior to thrusting indicating a possible
653 hiatus. Vertical fractures F1 and F3, produced during unloading and subsequent shrinkage,
654 are partitioned by sub-horizontal fracture F2. These fractures could have developed broadly
655 contemporaneously with F2 acting to partition stress, restricting the spatial development of
656 F1 and F3. Alternatively, F1 could predate F2, and F3 post-date F2. This would lend further
657 support to the interpretation that F2 is superimposed upon a relict thrust plane and that a
658 hiatus occurred during the accretion of LFA 1 resulting in partial consolidation and drying.
659 These characteristics, coupled with the emplacement of LFA 2 over LFA 1, are considered
660 to suggest a highly-dynamic temperate ice-marginal landsystem (cf. Evans and Twigg,
661 2002), characterised by multiple ice-marginal oscillations that resulted in the thrust-stacking
662 of multiple till blocks.

663 Till sequences produced by thrust-stacking typically create vertical, repetitive
664 sedimentological/lithological signatures observed frequently in exposures of glacial geology
665 elsewhere along the east coast of Britain (Boston *et al.*, 2010; Evans and Thomson, 2010;
666 Lee *et al.*, 2013, 2017). Although there are undoubtedly two till units (LFA 1 and LFA 2)
667 present at Tunstall, they have both been deposited by the NSL and reflect repeated ice-
668 marginal oscillations and till emplacement by thrust-stacking. We propose that the sequence
669 at Tunstall was generated by successive ice-margin oscillations, and are thereby suggestive
670 of active retreat (Boulton, 1996a, b; cf. Dove *et al.*, 2018). Alternatively, as the substrate is

671 progressively buried by surficial deposits following each advance-retreat phase, the stacked
672 sequence could be a function of reduced interaction of basal ice with the local lithologies
673 (Boulton, 1996a, b; Kjær *et al.*, 2006). In this case, the frequency of far-travelled
674 lithological components of the till would increase with height and produce a similar
675 lithological heterogeneity within samples, such as the results of this study.

676 *6.3 A litho-tectonic event model for Tunstall*

677 The fracture sets, lithological and sedimentological data from Tunstall enable an integrated
678 litho-tectonic model for the site to be proposed (**Table 3**). In summary, the vertical and
679 horizontal fractures provide a complementary record to the sedimentological and
680 lithological analysis suggesting multiple phases of ice advance and retreat: (i) ice advance
681 and accretion of the lower part of LFA 1; (ii) ice-marginal retreat, unloading, consolidation
682 and shrinkage (formation of F1); (iii) brief hiatus; (iv) ice-marginal re-advance, thrust
683 stacking of LFA 1 on top of LFA 1; (v) ice marginal retreat, unloading and horizontal
684 fracturing (F2) developed along a pre-existing decollement surface; (vi) ice-marginal re-
685 advance, emplacement of LFA 2; (vii) ice-marginal retreat, unloading, consolidation and
686 shrinkage – formation of F3 fractures; (viii) sub-marginal deposition of glaciofluvial
687 deposits LFA 3; (ix) non-glacial deposition of LFA 4 (x) coastal erosion and development of
688 lateral release joints (formation of F4).

689 The application of lithostratigraphic principals to the Skipsea Till significantly under-
690 represents the geological relevance of the ‘unit’ and specifically the number of ice-advances
691 that formed the unit. This mirrors studies utilising glaciotectionic evidence elsewhere in
692 eastern England, which records considerably much more dynamic phase of ice-marginal
693 behaviour than shown by lithostratigraphic data alone (Lee and Phillips, 2008; Phillips *et*
694 *al.*, 2008; Lee *et al.*, 2013; Phillips and Lee, 2013; Lee *et al.*, 2017).

695 The presence of two distinctly different till units – the Skipsea and Withernsea tills, is also
696 questioned based on lithological data. Instead the till units are interpreted as thrust-induced
697 stacks of pre-existing till that accreted during cyclical oscillations of the ice margin (cf.
698 Hiemstra *et al.*, 2007; Lee *et al.*, 2017) and up-ice variations in subglacial entrainment of
699 bedrock lithologies. We therefore propose the term ‘Skipsea till complex’ to encompass the
700 multiple thrust-stacked layers of till which cannot be classified lithostratigraphically.

701

702 **7. CONCLUSIONS**

703 The stratigraphic succession at Tunstall contains four lithofacies associations (LFA 1 - 4)
704 and three primary fracture sets (F1 - F3) which collectively establish an event stratigraphy
705 for the site encompassing the Late Devensian glaciation. The succession records two
706 superimposed subglacial traction till units (LFA 1 and 2). Sub-horizontal fracture (F2) is
707 interpreted as an unloading joint superimposed upon a regionally-extensive décollement
708 surface formed during the thrust-stacking of a unit of LFA 1 on top of LFA 1. Vertical
709 fractures F1 and F3 appear to relate to shrinkage and are tentatively interpreted to indicate
710 two separate phases of unloading and shrinkage that occurred following ice-marginal retreat
711 intra-LFA 1 and following the accretion of LFA 2. LFA 3 and 4 record the transition to non-
712 glacial conditions.

713 Clast lithological data have also been used to reconstruct glacial transport pathways for the
714 tills at Tunstall during the Late Devensian glaciation. The data provide evidence in support
715 of deposition by a lobe of glacier ice (first sourced from southern and central Scotland), that
716 flowed southwards down the present east coast of England (and offshore area of the
717 southern North Sea) before reaching its final extent on the Holderness coast. The
718 provenance of both tills therefore indicate an exclusively northern British origin. However,

719 statistical analysis of the clast lithological data demonstrates that there is greater intra-till
720 lithological variability within the tills than between the till units. We suggest this
721 lithological variability reflects temporal and spatial variability in the availability and
722 entrainment of bedrock source materials along the ice flow path. Furthermore, the two
723 subglacial tills at Tunstall cannot be differentiated lithostratigraphically, nor can they be
724 directly correlated with the regional glacial lithostratigraphy along the east and northeast
725 coast of Britain. This supports similar assertions made previously by Boston *et al.* (2010)
726 based on geochemical analysis from the Holderness tills.

727 Collectively, this evidence demonstrates that caution is required when applying
728 lithostratigraphic principles to till because these can underestimate the history of ice
729 advance and dynamic ice-marginal behaviour. In the case of this study, we consider that
730 variations in till lithological properties are not clear-cut between till units. Instead, the
731 variability in till composition reflects the temporal and spatial patterns of up-ice source
732 material entrainment plus the local erosional processes driven by thrust-stacking at an
733 oscillating ice margin.

734

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745

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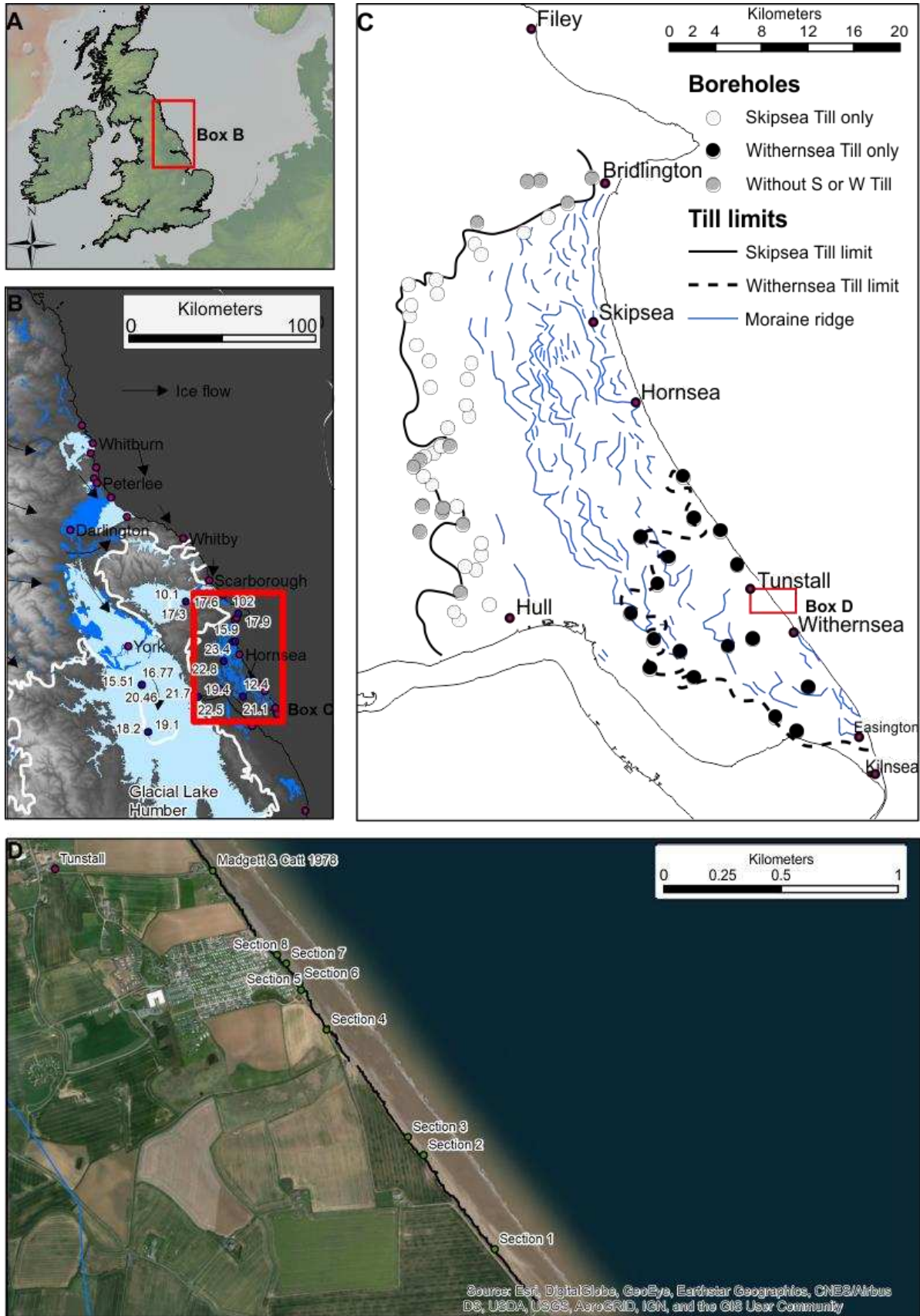
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1146 184

1147 **FIGURE CAPTIONS**



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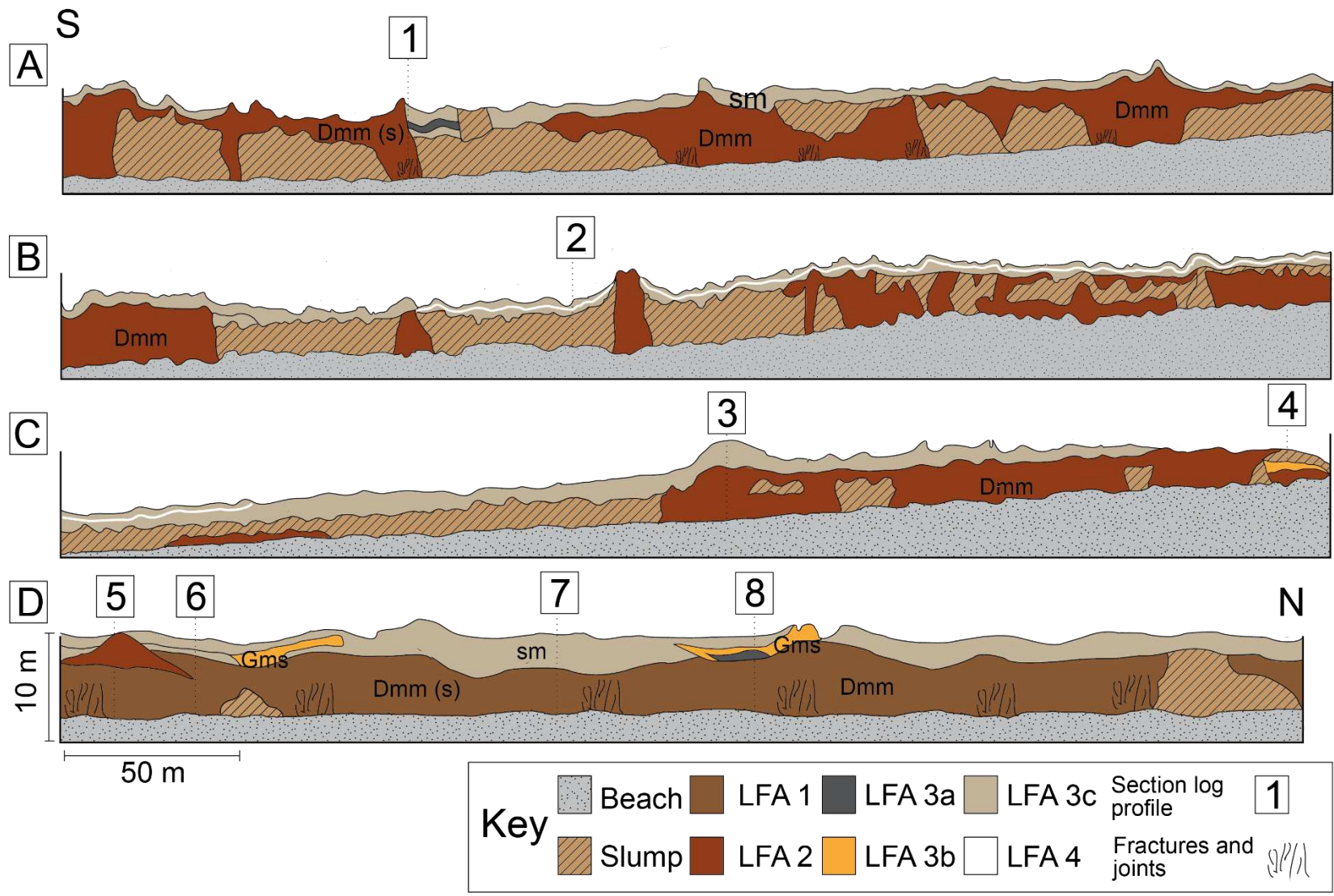
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Figure 1. A. Great Britain and study area highlighted. B. The Yorkshire and Durham coastline, with places named in the text. C. Study area, showing limits of Skipsea and Withernsea Tills (from Evans and Thomson, 2010). Published ages and geomorphology

1152 from Clark *et al.* (2018); Bateman *et al.* (2015; 2017); Evans *et al.* (2016). **D.** Detail of
1153 study area, showing location of section logs. Imagery from ArcMap Basemap
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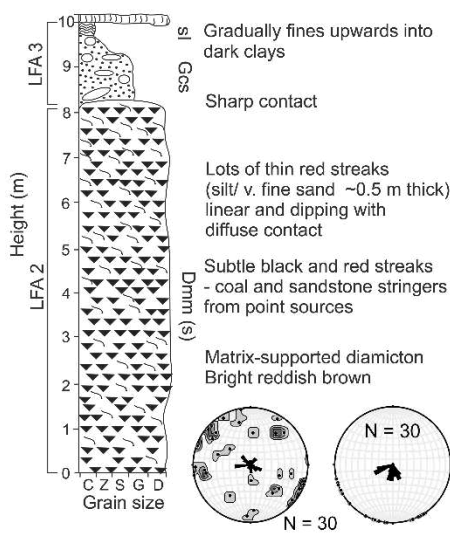


1156

1157 **Figure 2.** Facies architecture with clast fabric and strata orientation at Tunstall

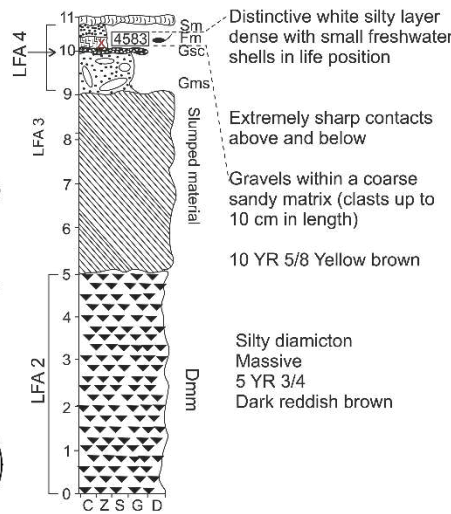
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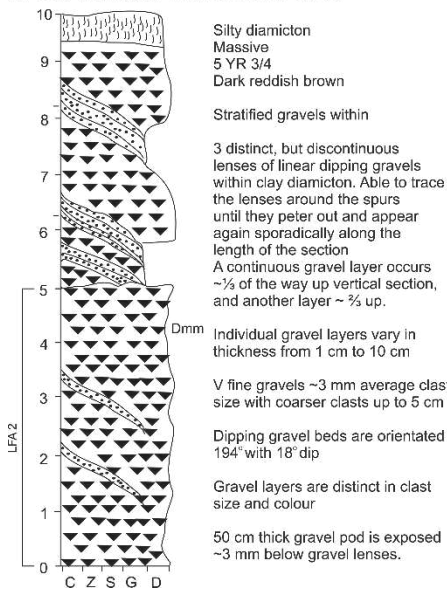
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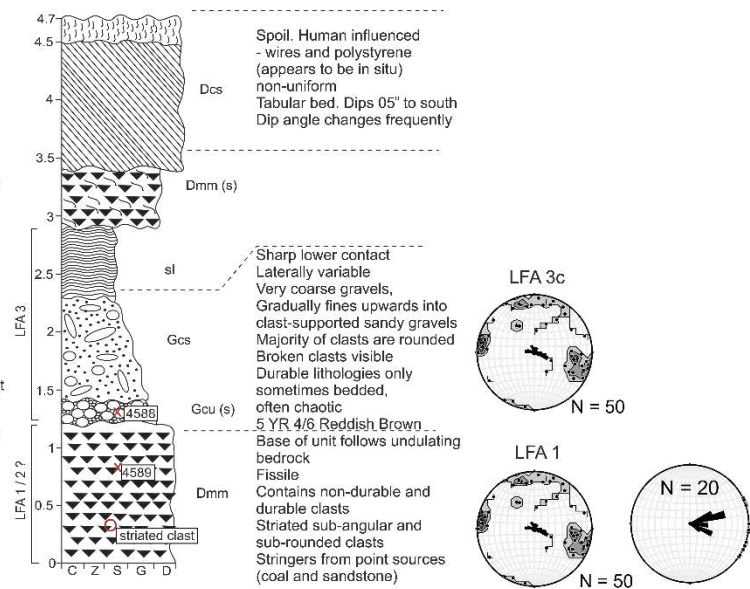
Section log no. 3

GPS co-ordinates : TA 31848 BNG 31109



Section log no. 4

GPS co-ordinates : TA 31869 BNG 31111



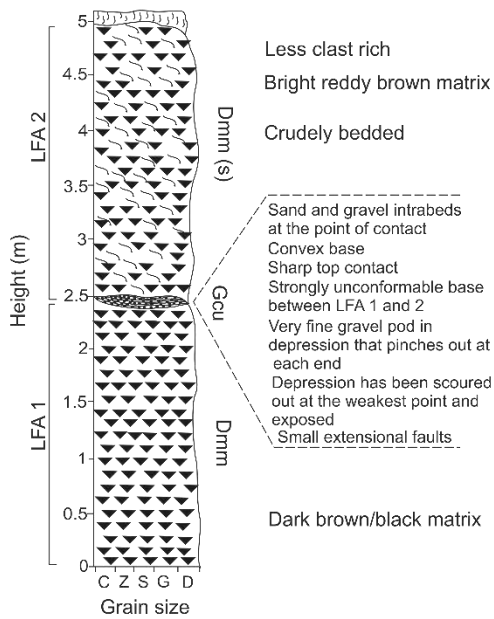
1158

1159 **Figure 3a.** Section logs 1 – 4 (see Figure 2 for log locations). Note differences in vertical

1160 scale

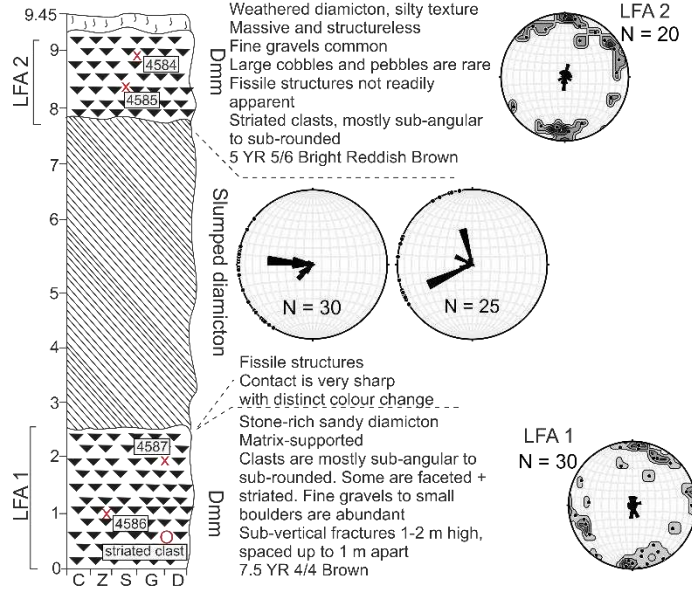
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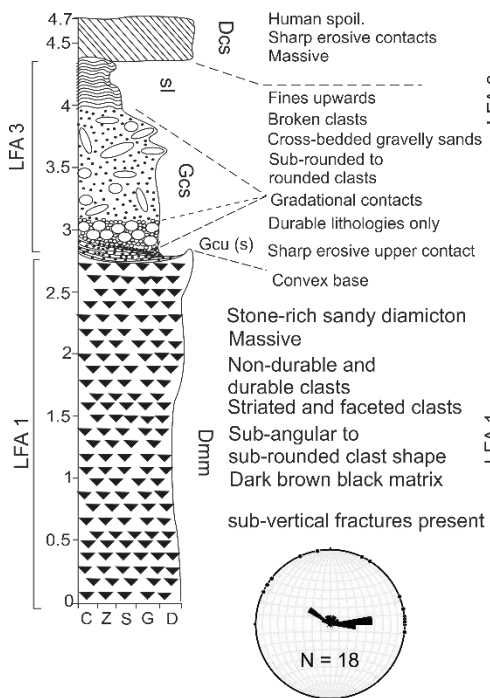
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GPS Co-ordinates : TR 31741 BNG 31283



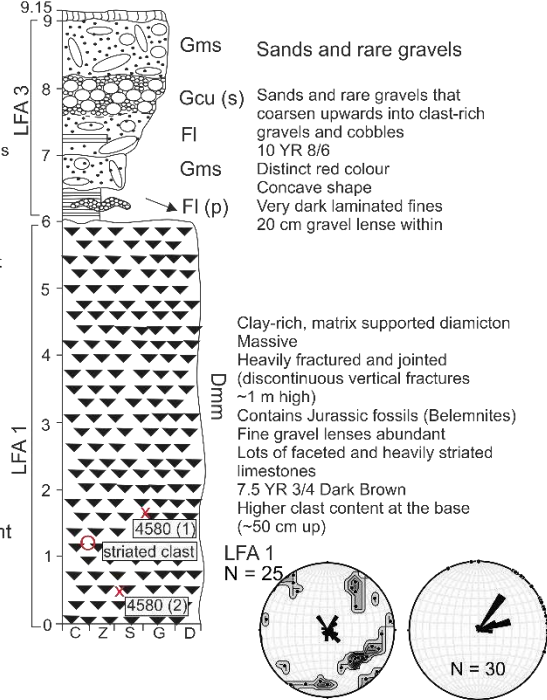
Section log no. 7

GPS co-ordinates : TA 31967 BNG 31120



Section log no. 8

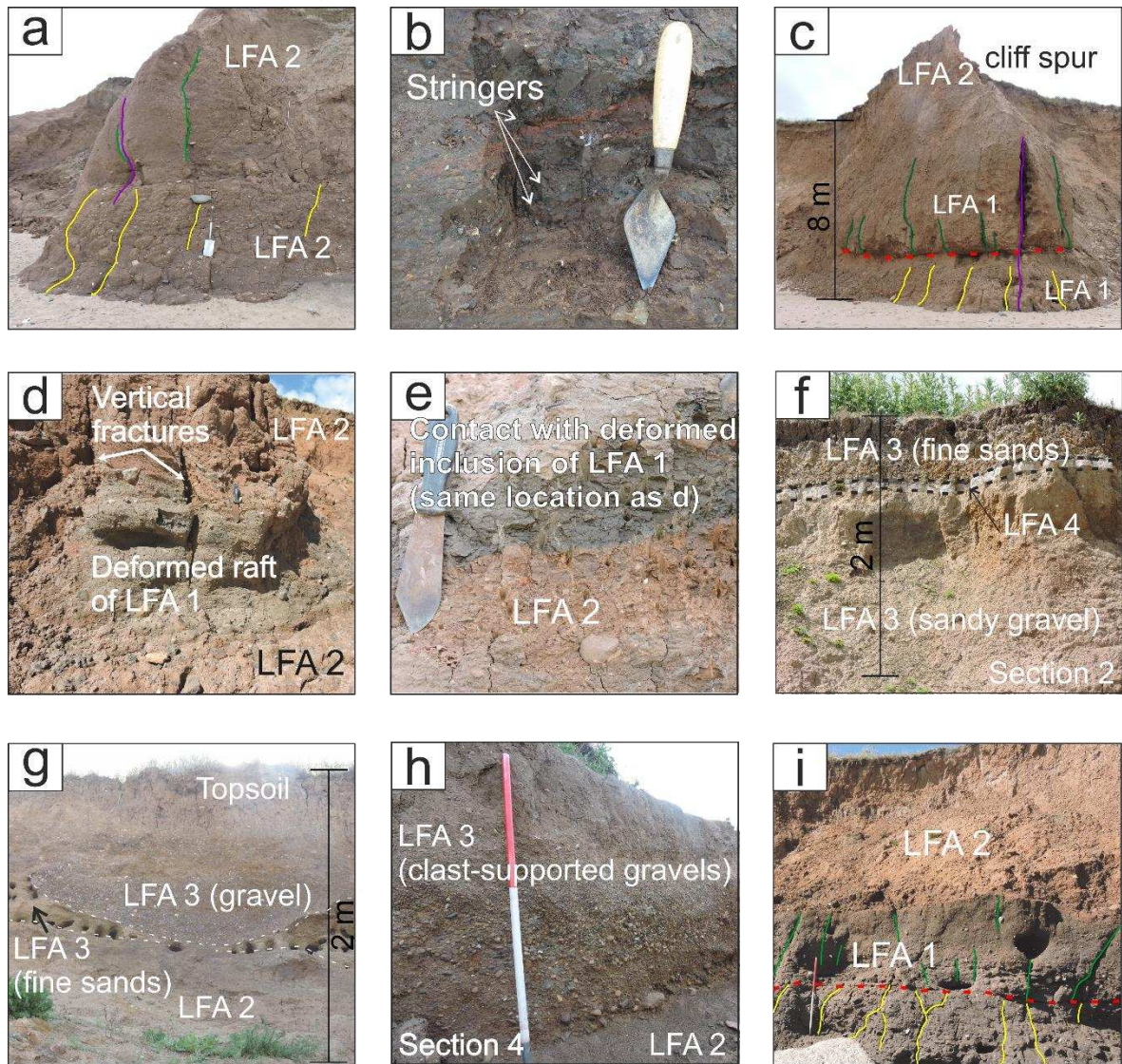
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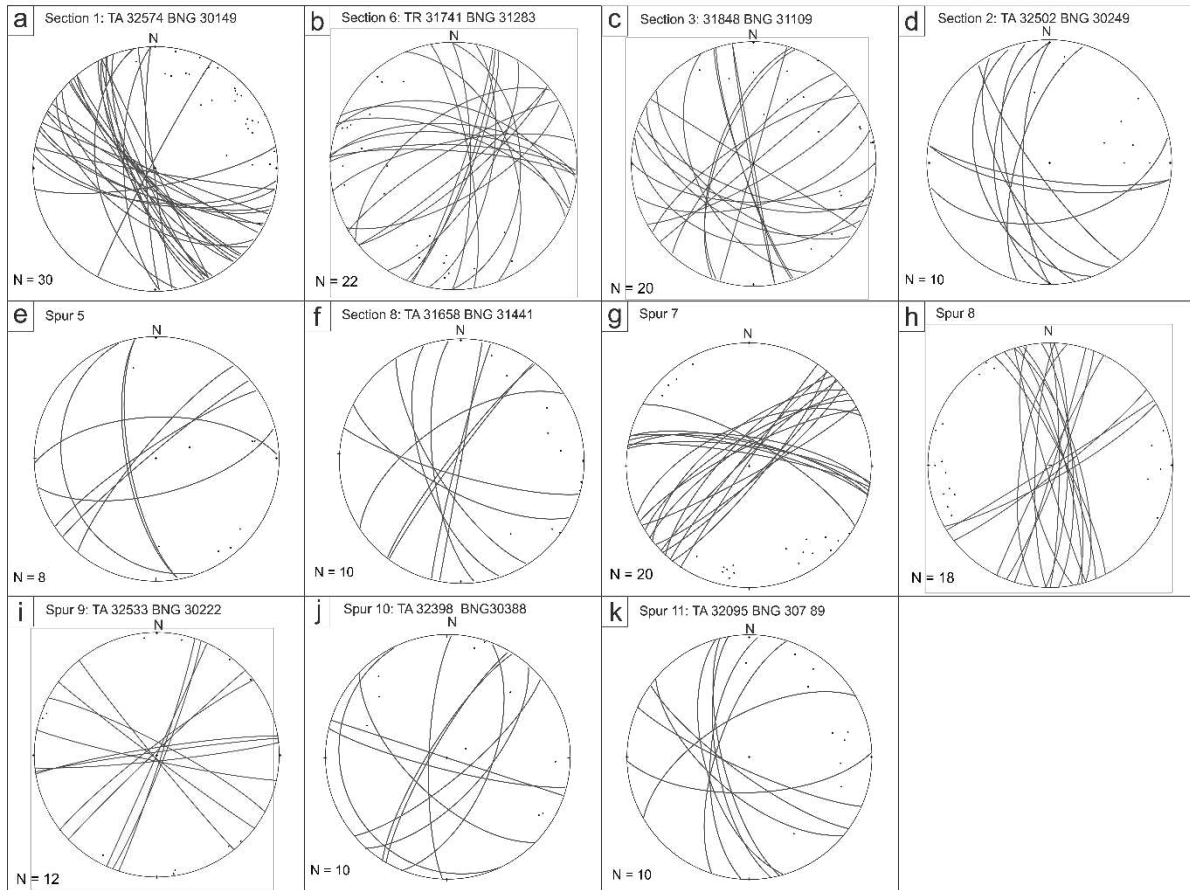
1162 **Figure 3b.** Section logs 5 – 8 (see Figure 2 for log locations). Note differences in vertical

1163 scale



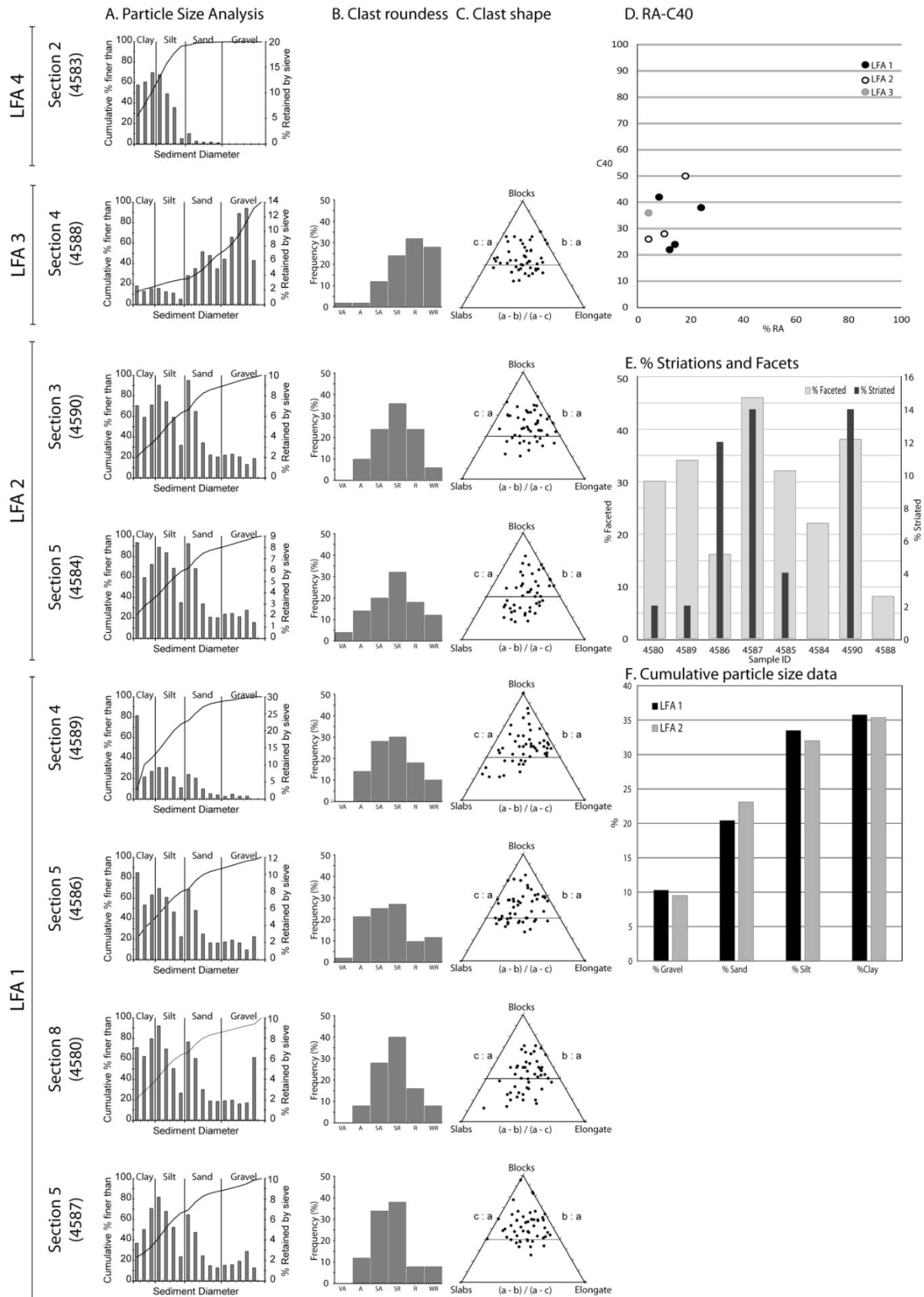
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1165 **Figure 4.** Representative photographs of key features at Tunstall including fracture sets F1
 1166 (yellow), F2 (red), F3 (green) and F4 (purple). **a.** LFA 1 and LFA 2 in superposition
 1167 showing fracture sets **b.** Stringers of red and black diamicton up to 10 cm thick in LFA 1 **c.**
 1168 Fracture sets F1, F2 and F3 in LFA 1 **d.** Raft of LFA 1 in LFA 2 **e.** Contact between LFA 2
 1169 and the deformed inclusion of LFA 1 **f.** Nature of LFA 4 **g.** Channel structure **h.** Cast-
 1170 supported gravels in LFA 3 **i.** LFA 1 and 2 in superposition, showing F1 and F2 fracture sets



1171

1172 **Figure 5.** Fracture measurements (dip angle and dip azimuth) plotted as poles to planes and
 1173 great circles on stereographic projections.

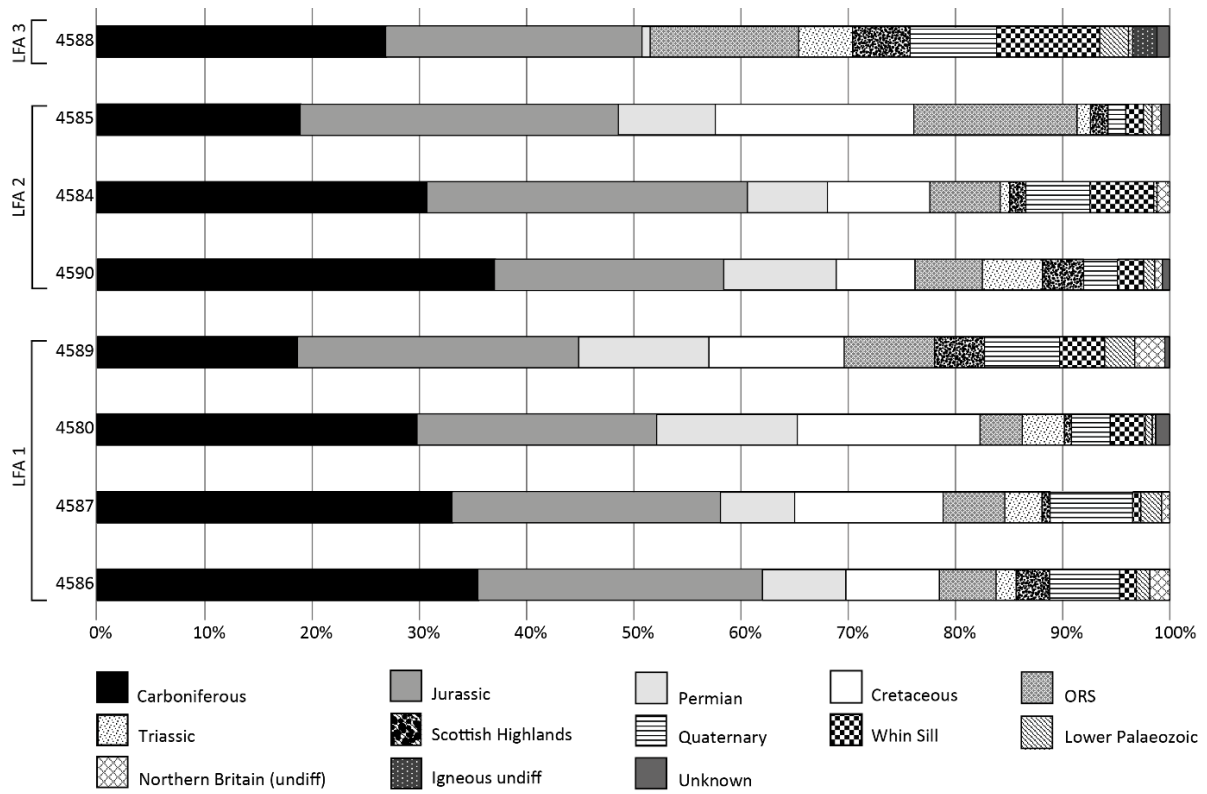


1174

1175 **Figure 6. A.** Particle size distribution for each sample. **B.** Clast roundness for gravel

1176 fraction. **C.** Clast shape for gravel fraction. **D.** RA-C40 graph. **E.** Percentage of striated and

1177 faceted stones within the gravel fraction. **F.** Total percentages of clay, silt, sand and gravel
 1178 in each lithofacies.



1179

1180 **Figure 7.** Clast lithological analysis for each sample



1181

1182 **Figure 8.** Representative photographs of key erratic lithologies in each sample **a.**

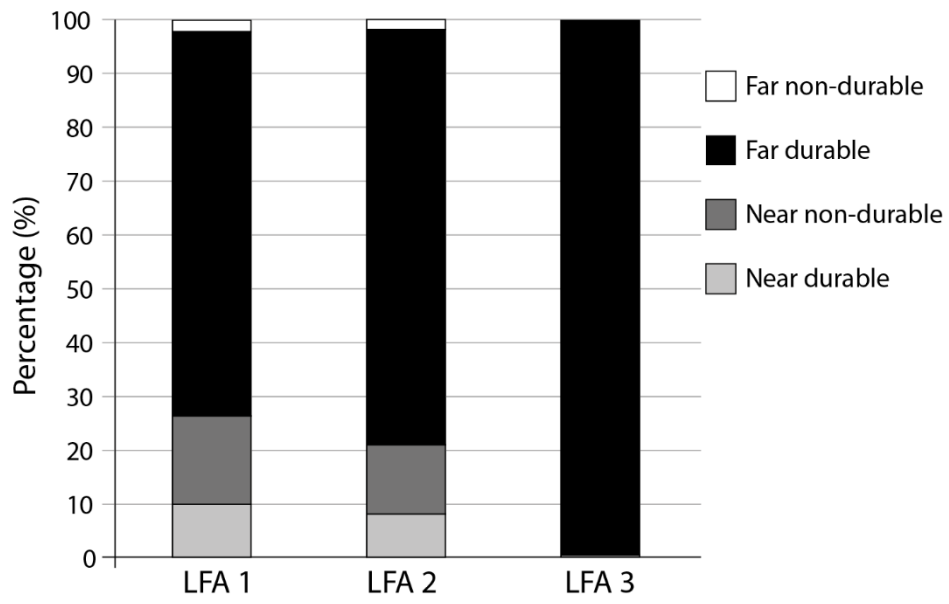
1183 Greywacke (LFA 2) **b.** Whin Sill Dolerite (LFA 2) **c.** Old Red Sandstone (LFA 1) **d.**

1184 Andesitic porphyry (LFA 2) **e.** Sherwood Sandstone (LFA 1) **f.** Magnesian Limestone (LFA

1185 1) **g.** K-feldspar rich Granite (LFA 2) **h.** Rhyolite (LFA 3) **i.** Carboniferous Limestone (LFA

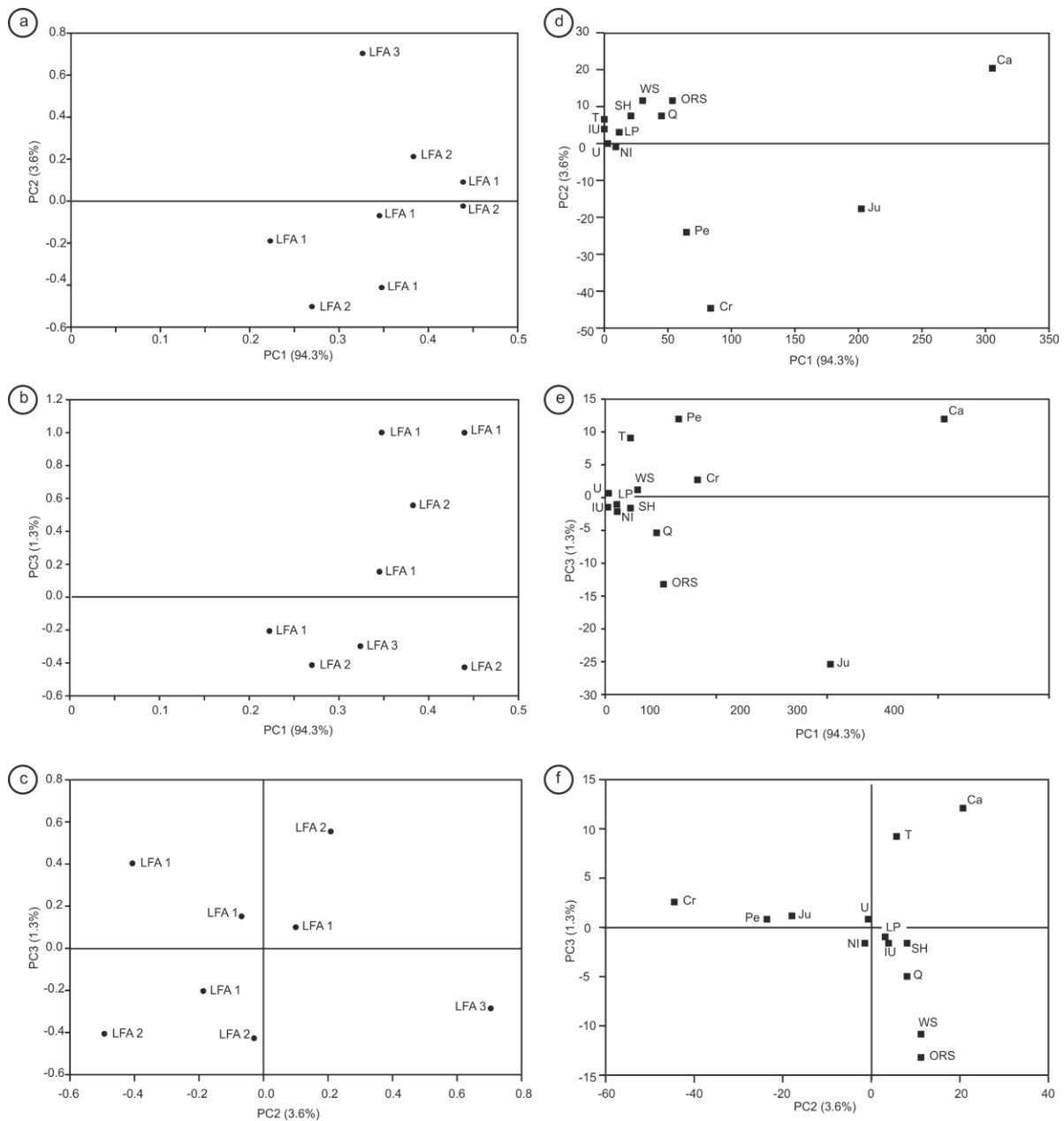
1186 1) **j.** K-feldspar < Quartz Granodiorite (LFA 1) **k.** Cheviot Granite (LFA 3) **l.** Quartzite

1187 (LFA 1).



1188

1189 **Figure 9.** Durability of far travelled and local clasts in each lithofacies



1190

1191 **Figure 10.** Plotting of processed PCA data. **a.** Principal component scores PC 1 and PC 2 **b.**

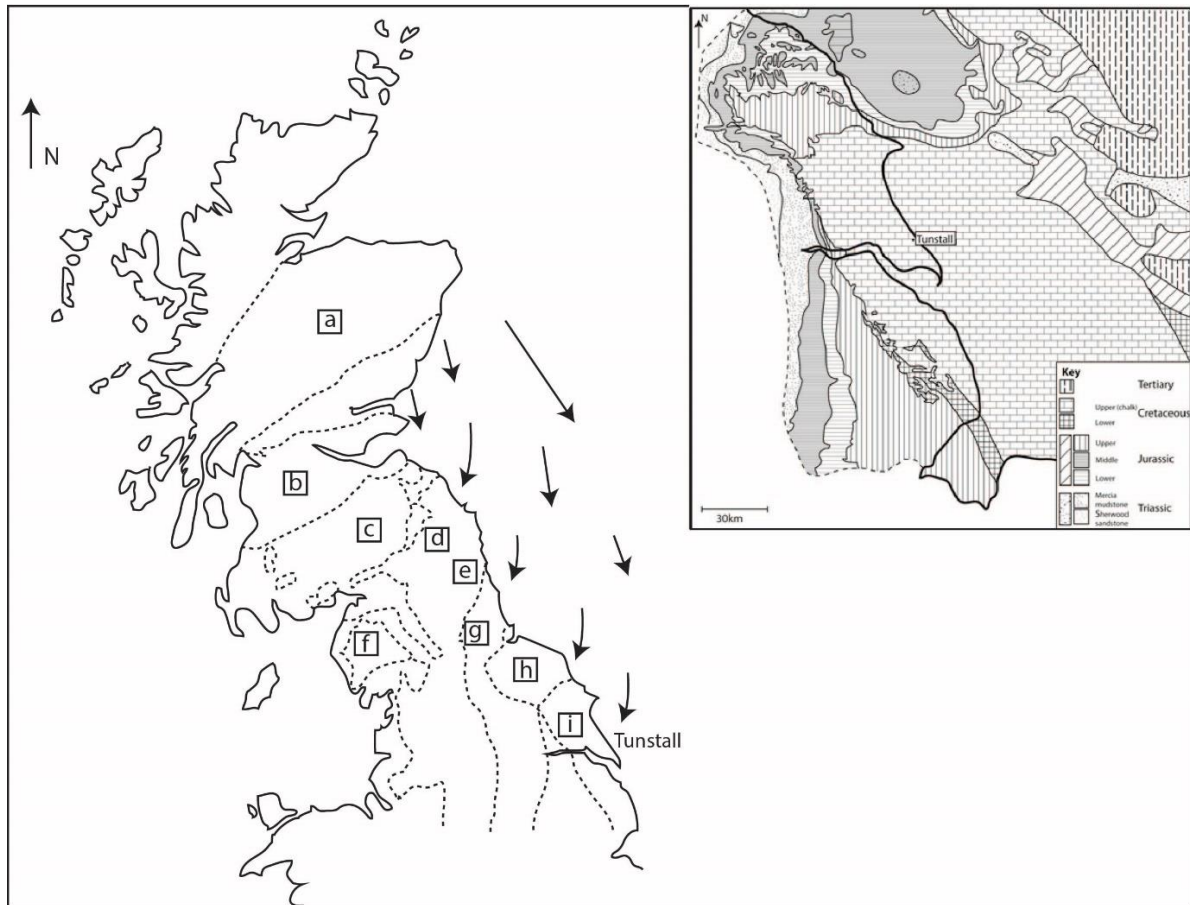
1192 Principal component scores PC 1 and PC 3 **c.** Principal component scores PC 2 and PC 3 **d.**

1193 Principal Component coefficients displaying lithological relationships between PC1 and

1194 PC2 **e.** Principal Component coefficients displaying lithological relationships between PC1

1195 and PC3 **f.** Principal Component coefficients displaying lithological relationships between

1196 PC2 and PC3



1197

1198 **Figure 11.** Revised iceflow pathways inferred from the simplified bedrock geology map of
 1199 Northern Britain with outcrop occurrences of the lithostratigraphical group .a. Grampian
 1200 Highlands b. Midland Valley c. southern uplands d. Cheviot volcanic complex e.
 1201 Northumberland f. Lake District volcanic complex g. County Durham h. Cleveland basin i.
 1202 Yorkshire basin Insert – Detailed map of the solid geology from the Tees estuary to The
 1203 Wash (adapted from Kent and Gaunt, 1980; Busfield *et al.*, 2015).

1204 **TABLE CAPTIONS**

	LFA 1	LFA 2	LFA 3	LFA 4
% Gravel	10.29	9.5	47.07	0.00
% Sand	20.41	23.1	27.74	3.22
% Silt	33.47	32	8.44	45.29
% Clay	35.78	35.4	16.75	51.49

1205 **Table 1.** Particle size analysis for each lithofacies at Tunstall

1206

1207

	Clast Lithology	LFA 1	LFA 2	LFA 3
Total (n)		893	847	260
Carboniferous	Arkose sandstone	2.46	4.57	0.00
	Carboniferous sandstone	13.66	18.44	26.54
	Carboniferous Chert	0.00	0.16	0.00
	Carboniferous Mudstone	0.00	0.16	0.00
	Carboniferous Limestone	9.52	7.38	0.00
	Coal	2.13	1.44	0.00
	Limestone undiff	3.81	0.64	0.00
	Basaltic Porphyry	0.56	0.96	0.38
	Carboniferous Porphyry	0.45	0.00	0.00
Permian	Sherwood Sandstone	2.69	3.05	0.77
	Magnesian Limestone	6.61	5.77	0.00
Triassic	Mercia Mudstone	3.02	3.05	0.00
	Red Siltstone	0.00	0.00	5.00
Jurassic	Jurassic Sandstone	6.16	7.54	3.08
	Yellow Sandstone	1.34	1.28	5.38
	Quartzitic Sandstone	1.12	0.96	0.00
	Siltstone undiff	4.03	6.26	8.46
	Yellow Siltstone	0.22	0.16	0.00
	Jurassic Mudstone	11.65	6.98	6.54
	Ironstone	0.00	0.80	0.38
	Jurassic Limestone	0.56	1.76	0.00
	Oolitic Limestone	0.00	0.16	0.00
Cretaceous	Cretaceous Chert	0.00	0.16	0.00
	Chert undiff	2.58	1.92	0.00
	Chalk	10.41	6.58	0.00
	Flint	0.22	0.32	1.54
Quaternary	Silcrete	0.00	0.16	0.00
	Brown Quartzite/ Vein Quartz	3.25	2.41	4.23
	Red Quartzite/ Vein Quartz	0.56	0.48	0.00
	White Quartzite/Vein Quartz	1.79	1.28	2.31
Lower Palaeozoic	Greywacke	1.23	0.48	2.69
Northern Britain	Gabbro	0.00	0.32	0.38
	Basalt	1.01	0.64	0.00
Whin Sill	Whin Sill Dolerite	1.90	4.33	9.62
Old Red Sandstone	Old Red Sandstone	2.91	2.25	6.15
	Old Red Sandstone porphyry	0.34	0.48	0.00
Scottish Highlands	Diorite	0.78	0.32	1.92
	Grano-diorite	0.22	0.16	0.00
	Micro-granite	0.00	1.12	3.08
	Granite	0.00	0.00	0.38
	Schist	0.45	0.48	0.00
	Phyllite	0.22	0.32	0.00
	Gneiss	0.00	0.08	0.00
Igneous	Igneous undiff	0.00	0.00	2.31
	Rhyolite	0.45	1.12	1.15
	Andesite	0.56	0.96	1.54
	Felsite	0.00	0.32	0.00
	Andesitic Porphyry	0.45	0.64	1.54
	Rhyolitic Porphyry	0.22	0.80	3.46
Unknown		0.45	0.32	1.15

1209 **Table 2.** Clast lithological data, showing percentages of clasts in each lithofacies

Event/Stage	Description	Interpretation	Implication
I	Massive, matrix-supported diamicton	Deposition of subglacial traction till (LFA 1; lower Skipsea Till)	Initial advance from NSL
II	Sub-vertical fractures (upwards to F2)	Unloading and shrinkage; development of F1 fractures	Sub-aerial exposure of ice-marginal retreat or thinning
Hiatus (Short)			
IV	Massive-matrix-supported diamicton	Thrust-stacking of LFA 1 (derived from nearby but up-ice) ontop of LFA 1 (décollement surface eventually became F2)	Re-advance of NSL
V	Sub-horizontal fracturing along pre-existing plane of weakness (décollement developed in stage iv)	Unloading and development of F2 fractures	Retreat of the NSL
VI	Massive, matrix-supported diamicton	Subglacial emplacement of LFA 2	Re-advance of NSL
VII	Sub-vertical fractures (upwards from F2)	Unloading and shrinkage; development of F3 fractures	Subaerial exposure of ice-marginal retreat or thinning
VIII	Sands and gravels	Deposition of LFA 3; proximal glaciofluvial outwash	Retreat following advance but no overriding of site; final stages of deglaciation
IX	White organic silt	Deposition of LFA 4 (Tufa)	Spring-fed pool under temperate icefree conditions

X

Sub-vertical
fractures (F4)

Lateral release joints

Coastal erosion

1211 **Table 3.** Summary of litho-tectonic event stratigraphy described from the base (oldest) to
1212 the top (youngest) in superpositional order

1213 SUPPLEMENTARY INFORMATION

Code	Description
Diamictons	
<i>Dmm</i>	Matrix-supported, massive
<i>Dcs</i>	Clast-supported, massive
---(<i>s</i>)	Stratified
Gravels	
<i>Gms</i>	Matrix-supported, massive
<i>Gcs</i>	Clast-supported
<i>Gcu</i>	Upwards coarsening gravels (inverse grading)
---(<i>s</i>)	Stratified
Sands	
<i>Sm</i>	Massive
<i>Sl</i>	Horizontal and draped lamination
Silts and Clays	
<i>Fm</i>	Massive
<i>Fl</i>	Fine lamination (minor sand and very small

ripples

---(*p*)

Intraclast or lens

1214 **SUPPLEMENTARY Table A1.** Lithofacies codes used in this study (and those in **Figure**

1215 **3a, b**), adapted from Benn and Evans (1998)