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1 **Giant scour-fills in ancient channel-lobe transition zones: formative**
2 **processes and depositional architecture**

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13 **Abstract**

14 Scours are common features of modern deep-marine seascapes,
15 particularly downstream of the mouths of slope channels within
16 channel-lobe transition zones (CLTZs). Their dimensions can exceed
17 hundreds of metres in width and length, and tens of metres in
18 depth. However, the stratigraphic architecture of the infill of these
19 erosional bedforms is rarely described from the rock record and no
20 large (>100 m width) scours have been described in detail from
21 exhumed CLTZs. Here, the infill of two erosional features (0.5-1 km
22 long and 15-20 m thick) from the Permian Karoo Basin succession,
23 South Africa, are presented from palaeogeographically well-
24 constrained CLTZs; one from Fan 3 in the Tanqua depocentre and
25 one from Unit A5 in the Laingsburg depocentre. The basal erosion

26 surfaces of the features are asymmetric with steep, undulating, and
27 composite upstream margins, and low gradient simple downstream
28 margins. The basal infill consists of thin-bedded siltstone and
29 sandstone beds cut by closely-spaced scours; these beds are
30 interpreted as partially reworked fine grained tails of bypassing
31 flows with evidence for flow deflection. The erosional features are
32 interpreted as giant scour-fills. The internal architecture suggests
33 different evolutionary histories for each case. The Unit A5 scour-fill
34 shows a simple cut-and-fill history with lateral and upward
35 transitions from siltstone- to sandstone-prone deposits. In contrast,
36 the Fan 3 scour-fill shows headward erosion and lengthening of the
37 scour surface suggesting temporal changes in the interaction
38 between turbidity currents and the scour surface. This relationship
39 could support the occurrence of a hydraulic jump during part of the
40 fill history, while the majority of the fill represents deposition from
41 subcritical flows. Diversity in scour preservation mechanisms could
42 explain the variety in depositional histories. The architecture,
43 sedimentary facies and palaeoflow patterns of the scour-fills are
44 distinctly different to well documented adjacent basin-floor
45 channel-fills at the same stratigraphic levels. The recognition of
46 scour-fills helps to constrain their sedimentological and stratigraphic
47 expression in the subsurface, and to improve our understanding of
48 the stratigraphic architecture of channel-lobe transition zones.

49 **Keywords**

50 channel-lobe transition; base-of-slope; giant scours; scour-fill;
51 bypass; facies characteristics; Karoo Basin

52

53 **1. Introduction**

54 Large scours are readily recognised erosional bedforms on modern
55 deep-marine seabeds (e.g., Palanques et al., 1995; Morris et al.,
56 1998; Wynn et al., 2002a; Bonnel et al., 2005; Fildani et al., 2006;
57 Macdonald et al., 2011a; Maier et al., 2011; Shaw et al., 2013;
58 Covault et al., 2014; Paul et al., 2014). Commonly, these scours are
59 concentrated within channel-lobe transition zones (CLTZs), a
60 relatively unconfined area dominated by sediment bypass that
61 separates the mouths of channel feeder systems from lobes (Mutti
62 and Normark, 1987, 1991; Kenyon et al., 1995; Wynn et al., 2002a).
63 Scours commonly form fields consisting of many individual and
64 coalesced scours (e.g., Wynn et al., 2002a; Macdonald et al., 2011a;
65 Shaw et al., 2013). The occurrence of scours is commonly
66 interpreted (Komar, 1971; Mutti and Normark, 1987, 1991; Garcia
67 and Parker, 1989; Garcia, 1993; Macdonald et al., 2011a; Ito et al.,
68 2014), and occasionally demonstrated (Sumner et al., 2013), to be
69 related to flows that have undergone a hydraulic jump
70 (transformation from supercritical to subcritical flow conditions),
71 triggered by changes in flow velocity and/or density. These changes
72 in flow behaviour are predicted to occur in base-of-slope to basin
73 floor transitions where there are abrupt change in gradient and
74 degree of confinement (e.g. Alexander et al., 2008; Ito et al., 2008).
75 While observations of small-scale scours and megafaults in ancient
76 systems are abundant (e.g. Macdonald et al., 2011a), large-scale

77 features are not well documented. Megascours associated with
78 Mass Transport Deposits (MTDs) have been constrained by various
79 seismic examples (e.g., Moscardelli, 2006; Sawyer et al., 2009; Ortiz-
80 Karpf et al., 2015) on slope settings and in some outcrop examples
81 from lower slope to base-of-slope deposits (Pickering and Hilton,
82 1998, their Fig.63; Lee et al., 2004; Dakin et al., 2012). In these
83 cases, erosional depressions are tens of metres deep and filled with
84 chaotic deposits. In contrast, large scour-fills in turbidite systems are
85 rarely identified in outcrop, therefore their evolution and
86 architecture are poorly constrained. Dimensions of turbidite-filled
87 scours reported from outcrop-related studies include the Ross
88 Formation (Ireland) with typical dimensions of 0.3-3.5 m in depth
89 and 1 to 45 m in length (Chapin et al., 1994; Elliott, 2000a, 2000b;
90 Lien et al., 2003; Macdonald et al., 2011b), the Albian Black Flysch
91 (Spain) with 1-5 m deep 5-50 m wide scours (Vicente-Bravo and
92 Robles, 1995), the Cerro Toro Formation (Chile) with scour depths of
93 metres and widths of tens of metres (Winn and Dott, 1979; Jobe et
94 al., 2009) and the Windermere Group with scours up to several
95 decimetres deep and several tens of centimetres to many tens of
96 metres wide (Terlaky et al., 2015); composite scours up to several
97 metres deep are reported in the Macigno Costiero Fm., Italy
98 (Eggenhuisen et al., 2011) and the Boso Pensinsula (Japan) with
99 erosional features filled with backset bedding up to 140 m wide and
100 10 m deep (Ito et al., 2014). These dimensions are an order of
101 magnitude smaller than the scour dimensions described from
102 modern systems (> 10 m depth and > 100 m width) (e.g. Wynn et al.,

103 2002a; Macdonald et al., 2011a). Scour-fills may be
104 underrepresented in the rock record because outcrop limitations
105 mean that they may have been misidentified as channel-fills due to
106 cross-sectional similarity (Mutti and Normark, 1987, 1991; Wynn et
107 al., 2002a; Normark et al., 2009). Furthermore, the stratigraphic
108 expression of the CLTZ, including scour-fills, is rarely fully exposed or
109 well-constrained in ancient systems (Mutti and Normark, 1987,
110 1991; Gardner et al., 2003; Ito et al., 2014; van der Merwe et al.,
111 2014).

112 Here, the morphology and depositional architecture of two
113 exhumed large-scale erosional scours from the Permian succession
114 of the Karoo Basin, South Africa, are described in detail: one
115 example from Fan 3 of the Tanqua depocentre and the other from
116 Unit A within the Laingsburg depocentre. Previous mapping has
117 constrained the palaeogeographic context of both locations to areas
118 where there is a down-dip architectural change from channel- to
119 lobe-dominated deposits (Morris et al., 2000; Van der Werff and
120 Johnson, 2003; Sixsmith et al., 2004; Hodgson et al., 2006; Jobe et
121 al., 2012; Pr lat and Hodgson, 2013). The presented examples
122 exhibit different sedimentological and architectural characteristics
123 compared to basin floor channel-fills in adjacent stratigraphy. The
124 objectives of this paper are to: i) evaluate the origin of these
125 distinctive erosional features, ii) compare the erosional and
126 depositional history to channel-fills, iii) develop recognition criteria
127 for scour-fills in outcrop, iv) discuss the role of erosional bedforms in
128 improving our understanding of the stratigraphic expression of

129 CLTZs within ancient submarine systems, and v) aid investigations
130 into the role of hydraulic jumps in deep-water bedform
131 development. Accurate recognition and description of large-scale
132 erosional architectural elements has important implications for the
133 robust application of outcrop studies to improve reservoir models
134 and reduce uncertainty in subsurface investigations.

135

136 **2. Regional Setting**

137 The Karoo Basin is one of a number of Late Palaeozoic to Mesozoic
138 basins that formed at the southern margin of Gondwana (De Wit
139 and Ransome, 1992; Veevers et al., 1994; López-Gamundi and
140 Rossello, 1998). The Karoo Basin has been interpreted traditionally
141 as a retroarc foreland basin with subsidence purely caused by the
142 loading of the Cape Fold Belt (e.g., Johnson, 1991; Cole, 1992;
143 Visser, 1993; Veevers et al., 1994; Catuneanu et al., 1998). More
144 recent interpretations suggest that subsidence during the Permian
145 was caused by dynamic topography effects due to subduction
146 (Tankard et al., 2009) in a pre-foreland basin stage. The southwest
147 Karoo Basin is subdivided into the Laingsburg and the Tanqua
148 depocentres (Fig. 1) of which the deepwater fill of both depocentres
149 is represented by the Ecca Group. The Ecca Group (Fig. 2) comprises
150 a 2 km-thick shallowing-upward succession from distal basin-floor
151 through submarine slope to shelf-edge and shelf deltaic settings
152 (Wickens, 1994; Flint et al., 2011).

153 **2.1 Tanqua depocentre**

154 This study focuses on part of Fan 3 of the Skoorsteenberg
155 Formation, which is one of four sand-rich basin-floor fan systems
156 (Fig. 2) (Bouma and Wickens, 1991, 1994; Wickens and Bouma,
157 2000; Johnson et al., 2001). Fan 3 is the most extensively studied fan
158 system of the Skoorsteenberg Formation, as it shows the most
159 complete outcrop extent (Hodgson et al., 2006). The Fan 3 study
160 area (Kleine Riet Fontein) is located in the southwestern corner of
161 the Fan 3 outcrop, which is the most updip location (Fig. 1,3A). An
162 integrated outcrop and research borehole dataset has established
163 the isopach thickness of Fan 3, and the relative spatial and temporal
164 distribution of sedimentary facies, architectural elements and
165 palaeocurrents (Johnson et al., 2001; Hodgson et al., 2006; Prélat et
166 al., 2009; Groenenberg et al., 2010). The axis of the system is
167 located farther to the east along depositional strike in the
168 Ongeluksvier area (Fig. 3A) and is characterised by distributive
169 basin floor channel systems with overall palaeocurrent to the
170 north/north-east (Van der Werff and Johnson, 2003; Sullivan et al.,
171 2004; Hodgson et al., 2006; Luthi et al. 2006). The distributive
172 character of the channel-systems at Ongeluksvier (Fig. 3A), the
173 more deeply erosional character of the channels in overlying Fan 4
174 and Unit 5, and the thinning to the south (Oliveira et al., 2009), all
175 suggest that the southwestern outcrop-limit of Fan 3 is a proximal
176 off-axis base-of-slope setting (Johnson et al., 2001, Van der
177 Werff and Johnson, 2003; Luthi et al., 2006; Hodgson et al., 2006;
178 Jobe et al., 2012). The Kleine Riet Fontein area was previously

179 studied in detail by Jobe et al. (2012) and interpreted as an area
180 receiving unconfined flows, supported by the wide spatial
181 distribution of numerous metre-scale scour features.

182 **2.2 Laingsburg depocentre**

183 The proximal basin floor system of the Laingsburg Formation is
184 divided into Units A and B (Sixsmith et al., 2004; Brunt et al., 2013)
185 (Fig.2). The 350 m thick Unit A comprises sand-prone sub-units A1-
186 A7, which are separated by regionally extensive mudstones
187 (Sixsmith et al., 2004; Flint et al., 2011; Pr lat and Hodgson, 2013).
188 The studied outcrop is in the 'Wilgerhout' area within Unit A5, a 100
189 m thick and tens of km-long package of sandstones and siltstones on
190 the northern limb of the post-depositional Baviaans syncline (Figs. 1,
191 3B), close to the town of Laingsburg. Palaeogeographically, the
192 study area is located in the axis of the A5 system on the basin floor
193 (Sixsmith et al. 2004) (Fig.1). The large number of sand-rich channel-
194 fills that characterise the upper part of A5 in this area point to a
195 location close to the base-of-slope and/or close to the mouths of
196 large distributary channels (Sixsmith et al., 2004; Pr lat and
197 Hodgson, 2013).

198

199 **3. Methodology and datasets**

200 Stratigraphic correlations were completed in the field using closely-
201 spaced sedimentary logs, photomontages, and walking out key
202 surfaces with a handheld GPS to construct architectural panels. In

203 the Fan 3 KRF study area (4.6 km²), a total of 20 sedimentary logs
204 was collected (Fig. 3A). More than 550 palaeocurrent
205 measurements, primarily from ripple cross-lamination, were
206 collected and tied to specific stratigraphic units. Due to the inherent
207 variability in direction of ripple cross-laminations, a high number of
208 measurements was needed and only surfaces with a clear dominant
209 orientation were measured. The dip direction of ripple foresets was
210 measured where possible to ensure an accurate palaeoflow
211 direction. The main outcrop face consists of a 3.5 km long N-S
212 depositional dip section. Several E-W orientated gullies to the east
213 of the main outcrop face provide additional depositional strike
214 control (Fig. 3A). Thin siltstone packages within the regional
215 claystones between Fan 2 and Fan 3 provide local correlation
216 datums.

217 Within the Wilgerhout area of Unit A5 a total of 17 sedimentary logs
218 along a ~ 500m depositional dip (W-E) section was collected (Fig.
219 3B). The regional A5 to A6 mudstone (Sixsmith et al., 2004; Flint et
220 al., 2011; Cobain et al., 2015) was used as the datum for all
221 correlations. A total of 44 palaeocurrents was measured solely from
222 ripple cross-laminations and give an average eastward directed
223 palaeoflow (082°). Where the tectonic tilt was significant(> 20°) the
224 azimuth of well exposed planar foresets was measured and
225 restored.

226 Facies associations have been determined based on previous Karoo
227 Basin studies (Johnson et al., 2001; Van der Werff and Johnson

228 2003; Grecula et al., 2003; Hodgson et al., 2006; Brunt et al., 2013)
229 and extensive study of the Fan 3 and Unit A system. Bounding and
230 erosion surfaces have been identified with the help of bed
231 truncation and the absence of normal-grading structures at bed
232 tops.

233 **4. Facies associations**

234 The deep-water deposits of the Karoo Basin show a limited grain
235 size distribution ranging from claystone to fine-grained sandstones.
236 Both Fan 3 and Unit A consist of mainly thin-bedded siltstones and
237 very fine- to fine-grained sandstones. Flow conditions were
238 interpreted from the described facies characteristics. A total of six
239 distinct facies associations was identified based on field
240 observations and are described in detail below.

241 **4.1 Thick structureless sandstones (Fa1)**

242 Thick (>1 m) fine-grained sandstone beds with little to no internal
243 structure can form amalgamated sandstone packages up to 5 m
244 thick and tens to hundreds metres wide/long, that are commonly
245 tabular. Weak normal grading is observed at bed tops, where
246 planar- and ripple-cross lamination may be preserved. Locally, bed
247 bases and/or tops can show planar laminations and/or banding
248 (alternating lighter and darker bands, see section 4.3). The
249 sandstone beds can contain a minor amount of dispersed sub-
250 angular mudclasts (Fig. 4A). Flame structures and tool (drag) marks
251 are observed at bed bases.

252 These deposits are interpreted as rapid fall-out from sand rich high-
253 density turbidity currents (Kneller and Branney, 1995; Stow and
254 Johansson, 2002) with clasts representing traction-transported
255 bedload (See section 4.2 and 4.3 for interpretation of planar
256 laminated and banded intervals). Flame structures are associated
257 with syn-depositional dewatering (Stow and Johansson, 2002).

258 **4.2 Medium-bedded laminated sandstones (Fa2)**

259 These medium- to thick-bedded (0.2 to 3 m thick), very fine-grained
260 to fine-grained, sandstones show various sedimentary structures.
261 Ripple lamination, in particular climbing ripple lamination, is
262 abundant (25-70% of all laminated sandstones), showing high angles
263 of climb with stoss-side preserved lamination (>45° on stoss-side
264 preserved laminae) (Fig. 4B). Some beds show a clear upward
265 increase in the angle of climb and proportion of stoss-side laminae
266 preservation. Where planar lamination is present it is commonly at
267 bed bases. Bases of thicker-bedded structured beds are sharp and
268 the basal part, in comparison to the remaining laminated part of the
269 bed, is commonly structureless. Bed tops always show abrupt
270 normal grading to fine siltstone. Bed geometries can show lateral
271 thickness variations on 10-s of metres scale.

272 High angles of climb and stoss-side preservation in ripple-laminated
273 sandstones are indicative of rapid unidirectional aggradation rates
274 (Jopling and Walker, 1968; Allen, 1973; Jobe et al., 2012; Morris et
275 al., 2014). When sedimentation rate exceeds the rate of erosion at
276 the ripple reattachment point, the stoss-side deposition is preserved

277 and aggradational bedforms develop (Allen, 1973). This style of
278 tractional deposition is attributed to rapid deceleration of the flow
279 and deposition from moderate-to low-concentration turbidity
280 currents (Allen, 1973; Jobe et al., 2012). The planar laminations
281 within the structured sandstones are interpreted to be deposited
282 under upper stage plane bed conditions (Allen, 1984; Talling et al.,
283 2012).

284 **4.3 Banded sandstones (Fa3)**

285 This facies association comprises medium- to thick-bedded fine-
286 grained sandstones (20 to 200 cm, on average 40 cm), with diffuse
287 laminae of over 1 cm thickness (Fig. 4C,5A2). This style of lamination
288 is characterised by an alternation between lighter and darker bands,
289 and is referred to as banded sandstones. Lighter bands are well
290 sorted and quartz-rich, whereas plant fragments and/or mudstone
291 clasts and micaceous materials are commonly found within the
292 poorly sorted darker bands. The banding is dominantly planar and
293 parallel to sub-parallel, but can be mildly wavy. Internal cm-scale
294 scour surfaces are common, as is loading at the bases of lighter
295 bands. Within thicker beds dominated by banded sandstone beds,
296 the bases are structureless and sharp.

297 Banded sandstones differ from planar-laminated sandstones due to
298 the thickness of the laminae (>1 cm), the thickness of the laminated
299 interval within individual event beds (>1 m) and the absence of any
300 major grain size differences between laminae. The observations
301 indicate highly concentrated, aggradational but fluctuating flow

302 conditions. These conditions are present during deposition in
303 traction carpets under high-density turbidity currents (Lowe, 1982;
304 Sumner et al., 2008, 2012; Talling et al., 2012; Cartigny et al., 2013)
305 and have not been linked to a generic flow regime. This is
306 comparable to the H2 division of Haughton et al. (2009) and the
307 Type 2 tractional structures of Ito et al. (2014). The internal
308 truncations that have been observed can be explained by quasi-
309 steady behaviour within a stratified flow, however the extent and
310 size of these truncations (up to m-scale) support more drastically
311 waxing and waning flow behaviour. Combined with the thickness of
312 individual beds within this facies group, these deposits support an
313 interpretation of high aggradation rate and/or possibly long-
314 duration of individual events.

315 **4.4 Thin-bedded sandstones and siltstones (Fa4)**

316 Thin (<20 cm) very fine-grained sandstones are interbedded with
317 laminated siltstones (<1 cm to 5 cm). Ripple lamination, including
318 low-angle climbing ripple lamination, is common within the
319 sandstone beds. This facies group can be subdivided into 1) tabular
320 sandstones with planar and ripple laminations (Fa4-1), and 2)
321 lenticular sandstones and siltstones associated with numerous
322 centimetre-scale erosion surfaces (Fa4-2; Figs. 4D, 5A1,5A4). Locally,
323 Fa4-2 sandstone beds contain mudstone clasts (<1 cm) (Fig. 5A3)
324 and can be associated with mudstone and siltstone clast
325 conglomerates (max 0.5 m thick, 1-2 m long) that are clast-
326 supported with a fine-grained sandstone matrix (Fig. 5B3). Within

327 Unit A5, thin (<5 cm) medium-grained, poorly sorted lenticular
328 sandstone beds that are at least 10 m long (Fig. 5B1) are associated
329 with Fa4-2 siltstones. The Fa4-2 siltstones of Unit A5 are thicker
330 bedded (>3 cm) (Fig. 5B1).

331 The tabular bed geometry and predominance of current ripple
332 lamination in Fa 4-1 are interpreted to indicate deposition from
333 lower phase flow conditions within sluggish dilute turbidity currents
334 (e.g., Allen, 1984). The occasional planar laminated sandstone
335 indicates upper phase flow conditions (Best and Bridge, 1992) but
336 with a transition to lower phase flow conditions due to ripple-
337 laminated tops.

338 The Fa 4-2 group supports a higher energy environment compared
339 to the FA4-1 group, which is interpreted based on the local presence
340 of mudstone clasts and the numerous erosion surfaces. This facies
341 association is interpreted to represent a period dominated by
342 sediment bypass (e.g., Stevenson et al., 2015). The fine-grained
343 siltstone deposits of the Fa4-2 group are interpreted to represent
344 the tails of bypassing turbidity currents and continued reworking of
345 the substrate by long lived turbidity currents, similar to channel-
346 margin deposits (Grecula et al., 2003, Brunt et al., 2013). The
347 mudstone clast conglomerates are interpreted as bedload material,
348 derived from a mud-rich substrate, and therefore represent lag
349 deposits of highly energetic bypassing turbidity currents. Sediment
350 bypass zones can show fluctuations between depositional, erosional
351 and bypass processes when being around the erosion-deposition

352 threshold (e.g., Wynn et al., 2002a; Ito et al., 2014; Stevenson et al.,
353 2015). This facies association shares many similarities to the
354 sediment bypass facies identified within the CLTZ's of sand-detached
355 lobe systems in the Laingsburg area (van der Merwe et al., 2014).
356 These fluctuations might result in the interbedding of siltstones and
357 sandstones, including thin and lenticular medium-grained sandstone
358 beds, and multiple erosion surfaces. The presence of unusually thick
359 siltstone beds, and medium-grained sandstones, which are very rare
360 in the Ecca Group, within the Unit A5 Fa4-2 facies group is evidence
361 of localised deposition. The climbing ripple lamination within thin-
362 bedded sandstones indicates rapid aggradation rates (Allen, 1973;
363 Jobe et al., 2012).

364 **4.5 Soft sediment deformed (SSD) deposits (Fa5) and claystones** 365 **(Fa6)**

366 The Fa5 facies group is represented by localised tightly folded and
367 contorted heterolithic units (0.2-0.5 m thick) of thin-bedded
368 siltstones and sandstones (Fig. 5B2). Fa5 represents a minor portion
369 of the infill (<1%) and occurs only within the basal infill towards the
370 margins of both features, and in close association with Fa4-2, rarely
371 exceeding 2 m in length. Thick, regionally extensive units of Fa6
372 claystones occur as drapes to the deepwater sandstone units in both
373 depocetres.

374 Due to their marginal location and limited proportions, the
375 contorted thin bed units are interpreted to represent local
376 remobilisations, above erosional relief. Fa 6 represents condensed

377 intervals of hemipelagic deposition, during periods of regional
378 shutdown in coarse-grained sediment supply (Hodgson et al., 2006;
379 Luthi et al., 2006; Flint et al., 2011; van der Merwe et al., 2014).

380

381 **5. Depositional Architecture**

382 Both features are defined by composite and asymmetric basal
383 erosion surfaces that exceed the extent of the exposures (>350 m
384 long in A5; >1000 m long in Fan 3) and incise 15-20 m into
385 underlying deposits. The consistent palaeoflow directions of
386 underlying and overlying deposits indicate that the updip margins
387 are highly irregular, undulating and orientated sub-parallel to
388 regional palaeoflow directions. The Fan 3 exposure is orientated
389 150-330° with a 340° average palaeoflow at this location (n = 435).
390 The Unit A5 exposure is orientated 075-255° with a 082° average
391 palaeoflow (n = 44). The type and distribution of sedimentary facies
392 and internal stacking patterns differ between the two cases, and are
393 discussed separately.

394 **5.1 Fan 3 feature; Tanqua depocentre**

395 The location of the Fan 3 erosional feature (Fig. 6) is in the middle of
396 a north-south orientated outcrop of proximal off-axis deposits in the
397 Kleine Riet Fontein area (Figs. 1, 6). Mapping of thickness, facies,
398 and system-scale sedimentary architecture with the lack of channel-
399 fills in this area compared to the Ongeluks river area to the south-
400 east supports this as a proximal off-axis environment (Johnson et al.,

2001; Hodgson et al., 2006; Jobe et al., 2012) (Fig.3A). The overall
palaeoflow is northwards (Fig. 7). The underlying deposits can be
subdivided into a sandstone-prone package dominated by Fa2 and
an overlying siltstone-prone package dominated by Fa4 (Figs. 6 and
7). A minor stratigraphic change in mean palaeoflow is identified
between these two packages, from NNE (033°) to NNW (336°) (Fig.
7). The feature in the Kleine Riet Fontein area shows an erosional
cut into the siltstone-prone package. All deposits within Fan 3 below
the basal erosion surface of the studied feature extend beyond the
study area and are therefore more laterally extensive . The basal
erosion surface forms a series of metre-scale steps on the steep
(max. 50°) updip southern margin (Figs. 5A5, 6). The full geometry of
the northern margin is obscured, but the overall thinning of the fill
suggests a low-angle confining surface (Fig. 6). Sedimentary sections
taken towards the east (Fig. 3A) of the main N-S profile (Fig. 6)
indicate eastward shallowing of the basal erosion surface and infill
directed perpendicular to regional palaeoflow (Fig. 8A).

The architecture of the fill is characterised by abrupt changes in bed
thickness and multiple truncation surfaces (Figs. 6, 9,10), and can be
subdivided into two distinct infill packages based on changes in
facies proportions and architecture across key bounding surfaces
(Fig. 6, 8B). The lower package is up to 6.5m thick and is subdivided
into two distinct elements (1 and 2) based on an internal truncation
surface and abrupt changes in bed thickness (Fig. 6). Both elements
comprise (Fig. 8B) thin-bedded siltstones and climbing ripple
laminated sandstones (Fa4-2) containing small (1-4 cm) mudstone-

427 chips and minor (20-40 cm thick) folded thin-bedded deposits (Fa5).
428 The lower package is only present in the northern part of the fill.
429 The upper package is up to 12 m thick and is subdivided into five
430 different infill elements (3-7) (Figs. 6, 9, 10) based on abrupt
431 stratigraphic facies transitions and internal erosion surfaces. The
432 upper package elements are predominantly medium- to thick-
433 bedded structured sandstones (Fig. 8B) with upward steepening
434 climbing ripple-lamination and increased stoss-side preservation
435 (Fa2) (Fig. 4B). Each of these elements, varying between 2 and 7.5 m
436 in thickness, shows lateral thickness and facies changes. Within the
437 individual elements there are no clear vertical trends, however the
438 combination of the upper and lower packages together forms a
439 stepped coarsening and thickening upward profile (Fig. 8B) within
440 the axis of the feature. The lower elements in the upper package (3,
441 4,5) are more laterally restricted (Figs. 9,10) and show more
442 substantial bed thickness variability compared to the upper
443 elements (6 and 7). Element 4, and to a lesser extent element 6,
444 thicken where the underlying elements thin (Fig. 6 and 9). Facies
445 within the upper package are uniform with a dominance of climbing-
446 ripple laminated sandstone (Fa2) and a transition to more thinly-
447 bedded deposits (Fa4) towards the margins of all elements (Fig. 6)
448 where not truncated. Elements 5 and 7 contain some banded
449 sandstones (Fa3) directly overlying the basal erosion surface at the
450 southern margin (Fig. 5A2), which show a northward facies
451 transition to structureless sandstone (Fa1). Elements 3, 4 and 5
452 (Figs. 9 and 10) show bedsets (3-7 m thick) that comprise four to five

453 metre-thick (0.5 to 2.0 m) dominantly climbing-ripple laminated
454 sandstone beds, which are interbedded with thin siltstones (<0.1 m).
455 They are thickest near the southern margin, and pinch or taper out
456 into thin siltstones (<5 cm) in a northward direction (over 50-150 m).
457 Successive pinchouts occur southward, such that the beds shingle
458 updip. Where normally graded sandstone beds thicken they
459 amalgamate, as can be seen in element 5 (Fig. 10B). Due to
460 accessibility issues, the exact orientation of the bedding is difficult
461 to measure directly, but outcrop sections of element 3 and 5 (Fig. 9
462 and 10) indicate a shallow southward (updip) depositional dip (a few
463 degrees). In addition, beds in element 3 dip upstream approximately
464 2-4° relative to the basal erosional surface (Fig. 9). Elements 6 and 7
465 do not preserve clear bed stacking patterns, although lateral
466 variability in bed thickness is observed.

467 The palaeoflow patterns within the erosional feature in the Kleine
468 Riet Fontein area are diverse and show lateral and stratigraphic
469 variations. The lower package preserves a dominant south-easterly
470 orientation (134°) within element 1 at its most southern limit, which
471 becomes more eastward (093°) within element 2 (Fig. 6, K7).
472 Several hundred metres to the north, element 2 has a dominant NNE
473 to NE (025° – K8, 035° – K9; Figs. 6 and 7) palaeoflow direction. In
474 the upper package, there is a general NE to SE trend (096°) except
475 for element 7, which shows an overall NE direction (028°) at its
476 northern limit and a more NNW direction (345°) near its
477 southernmost limit (Fig. 6). The thin-bedded deposits above the

478 upper package have a NNE palaeoflow (028°, n = 57), consistent
479 with the underlying deposits and the regional trend.

480 A second large-scale erosional feature 800 m to the north (around
481 K13, Fig. 3A) is situated at the same stratigraphic level in Fan 3, and
482 shares many similarities in architecture and infill facies. The infill of
483 this erosive feature exhibits an average NNW-directed (330°) (n=17)
484 palaeoflow, similar to the underlying thin-bedded deposits. An
485 irregular erosion surface (~15°) has a measured ENE orientation
486 (070°), perpendicular to palaeoflow and overlain by structured
487 sandstones. Three hundred metres to the east, another erosion
488 surface (~6°) is approximately orientated NNW (335°), which is
489 parallel to the palaeoflow, and is evident from discordance in bed
490 dips within the thin-bedded deposits. The basal surface has a
491 minimum length of 500 m perpendicular to palaeoflow and cuts at
492 least 10 m into underlying deposits. Close to the western margin the
493 fill consists solely of Fa2 facies, but eastwards where the fill
494 thickens, it consists of a lower package of thin-bedded siltstones and
495 sandstones (Fa4-2) and an upper package of structured sandstones
496 (Fa2), that are locally amalgamated.

497 **5.2 Unit A5 feature; Laingsburg depocentre**

498 The erosional feature at Wilgerhout lies in the upper half of Unit A5
499 (60-70 m from the base) within a succession of stacked lobes locally
500 cut by sand-rich channel-forms (Prélat and Hodgson, 2013). A large
501 channel (>600 m wide and >15 m deep) filled by amalgamated
502 structureless sandstones (Fa1) is present at the top of the A5

503 succession in this location. The association of channels and lobes in
504 this area supports a base-of-slope setting, within the upper part of
505 the Unit A5 system based on regional mapping (Sixsmith et al., 2004;
506 Prélat and Hodgson, 2013). The exposure is limited to a 1 km long E-
507 W orientated section. Regional palaeoflow patterns are towards the
508 ENE (Sixsmith et al., 2004), which is consistent with measurements
509 from the infill deposits (Fig. 11). The section shows a steep (2-50°)
510 and stepped western (updip) margin.

511 The fill consists of three distinct sedimentary packages. A lower
512 package (1.5-5 m thick) comprises thin-bedded siltstones with rare
513 banded or ripple laminated fine-grained sandstone beds (Fa4-2) (Fig.
514 11). Locally, thin (<30 cm) mudstone clast conglomerates directly
515 overlie the basal erosion surface. Multiple small-scale (<20 cm deep)
516 cross-cutting erosional surfaces incise into thin-bedded siltstones in
517 the basal ~0.5 m of the fill, but decrease towards the top. Thin (2-3
518 cm) and lenticular moderately sorted medium-grained sandstones
519 are present within the siltstones and individual normally graded
520 siltstone beds are thick (> 3 cm) (Fig. 5B1). The middle package (0.5-
521 10 m thick) comprises medium-bedded banded (Fa3) and
522 structureless sandstones (Fa1) interbedded with siltstones, which
523 pass abruptly from sandstone-dominated to siltstone-dominated
524 (Fa4-2) at the western margin (Fig. 5B4). Some minor tightly folded
525 deposits (Fa5) (Fig. 5B2) occur within the siltstone dominated
526 western margin succession. The upper package (3-8.5 m thick)
527 comprises thick-bedded sharp-based structureless partially
528 amalgamated fine-grained sandstones (Fa1) interbedded with the

529 occasional banded sandstone and thin-bedded siltstone (Fig. 11).
530 This package extends beyond the limits of erosional confinement,
531 but increases in thickness above the deepest point of the basal
532 erosion surface. Within all three infill packages no clear vertical
533 stratigraphic trends have been observed. However, the combination
534 of the three infill packages together (Fig. 11B –W15) shows a step-
535 wise coarsening- and thickening-upward trend above the basal
536 surface. Above the upper package a 4 metre thick fining- and
537 thinning-upwards unit can be observed (Fig. 11) from thick- to
538 medium-bedded dominantly structureless sandstones (Fa1) to thin-
539 bedded sandstone and siltstone deposits (Fa 4). As these deposits
540 are tabular, they are not considered to be part of the fill. Up to 500
541 m west (updip) of the basal erosional surface at the same
542 stratigraphic level, a fine-grained thin-bedded package (1-2 m thick)
543 is characterised by multiple small-scale (10-20 cm) erosion surfaces,
544 thin sandstone beds, and abrupt bed thickness changes.

545

546 **6. Discussion**

547 **6.1 Origin and infill of erosional features**

548 The average palaeocurrents from underlying and overlying deposits
549 indicate that the large asymmetric erosion surfaces described from
550 Fan 3 (Tanqua) and Unit A5 (Laingsburg) are dip-sections (Figs. 3,
551 6,11). The steeper upstream surfaces dip at angles of 4-50° (Fan 3)
552 and 2-50° (Unit A5), with prominent metre-scale steps and multiple
553 erosion surfaces indicating the composite nature of the basal

554 surface. In the Kleine Riet Fontein area (Fan 3), the transverse to
555 downstream section is shallow ($\sim 3.5^\circ$) and smooth (Fig. 6), and
556 shows prominent asymmetry in three dimensions with shallowing of
557 the basal surface perpendicular to the regional palaeoflow (Fig.
558 10A). Palaeocurrents within the basal are diverse and at a high angle
559 (orientated N to SE) to underlying and overlying deposits (orientated
560 NNW), with palaeoflow differences of up to 180° (Fig. 7).

561 An asymmetric and composite basal erosional surface with stacked
562 smaller-scale elements could support an interpretation of a sinuous
563 submarine channel-fill. However, in a cut through a sinuous channel-
564 fill the palaeocurrents would be expected to be dominantly parallel
565 to channel banks (dip sections) (Parsons et al., 2010; Wei et al.,
566 2013; Sumner et al., 2014) and are only rarely described at high
567 angles to the basal surface (Pyles et al., 2012). Known exhumed
568 examples of outer bank deposits have relatively higher energy facies
569 such as conglomerates, and coarse-grained and/or amalgamated
570 sandstones (e.g., Young et al., 2003; Labourdette et al., 2007;
571 Hodgson et al., 2011; Janocko et al., 2013) compared to inner bank
572 deposits, which does not match with the observed distribution of
573 facies and the relatively low-energy character of the Kleine Riet
574 Fontein infill. Furthermore, channel sinuosity is predicted to be low
575 for the slope gradients of base-of-slope and basin-floor channel
576 bends (e.g., Clark et al., 1992 – close to 1.0 sinuosity at 1:1000 slope
577 angles), especially in sand-prone systems without levees and at mid-
578 high palaeolatitudes ($50\text{-}60^\circ\text{S}$) at which the Karoo system formed
579 (Peakall et al., 2012, 2013; Morris et al., 2014; Cossu et al., 2015).

580 Consequently, given the low predicted sinuosities it is unlikely that
581 channel bend facies would be at high angles to the regional slope. In
582 contrast, the morphology of the basal surface, the palaeocurrent
583 pattern, and distribution of sedimentary facies support an
584 interpretation of large-scale scour-fills, with prominent steep
585 headwalls and lower-angle downstream margins. In addition,
586 coarsening- and thickening-upwards trends as identified within both
587 features, considering the complete infill, are more readily explained
588 as a progradational trend (e.g., Macdonald et al., 2011b) than
589 channel-fills. Basin-floor channel-fills in the Ecca Group have been
590 described by several authors (e.g., Johnson et al., 2001; Sixsmith et
591 al., 2004; Sullivan et al., 2004; Brunt et al., 2013), and are
592 dominantly characterised by structureless sandstone, highly
593 amalgamated in the axis of the fills and more thin-bedded towards
594 the margins and top of the fills . Where well preserved, the basal
595 erosion surface and facies distribution of basin-floor channels are
596 symmetrical (Sullivan et al., 2004; Luthi et al., 2006), and typically
597 ~250-350 m wide and 15-20 m thick (Pringle et al., 2010; Brunt et
598 al., 2013). The large-scale scour-fills described, therefore, are
599 distinctly different to the published examples of basin-floor channel
600 fills from the Karoo Basin in terms of their architecture, facies types
601 and distributions, and relationship of palaeoflow to the bounding
602 surface, as well as to sinuous channels from other settings. The
603 erosional feature within the Kleine Riet Fontein area of Fan 3 has
604 been previously interpreted as a channel-fill. Morris et al. (2000)
605 classified it as a crevasse channel-fill, while Van der Werff and

606 Johnson (2003) interpreted it as the distal depositional part of an
607 overbank channel-fill with a SE-NW orientation. Implicit in both
608 interpretations was that the depositional architecture is different to
609 the basin-floor channel-fills at the same stratigraphic level 7-8 km to
610 the east (e.g., Sullivan et al., 2004; Luthi et al., 2006).

611 The basal fine-grained fill (Fa4-2) in both scour-fills is interpreted to
612 indicate sediment bypass and the deposition of low-energy tails of
613 flows. The interpretation of thin-bedded deposits indicating
614 sediment bypass has been previously made for channel-fills (e.g.,
615 Beaubouef and Friedmann, 2000; Grecula et al., 2003; Brunt et al.,
616 2013; Stevenson et al., 2015). However the thicknesses of individual
617 siltstone beds (>3 cm) within the Unit A5 fill is distinctive, and is
618 interpreted to reflect the capture of flow tails in a scour depression,
619 in a similar manner to thick siltstones in internal levee successions
620 (Kane and Hodgson, 2011). In the Kleine Riet Fontein feature (Fan 3),
621 the diverse palaeoflow directions in the basal siltstone units
622 suggests deflection and spreading of flow at the upstream end.
623 Complex flow patterns are known to be associated with flutes and
624 scours (e.g., Eggenhuisen et al., 2011) with flows exhibiting a
625 recirculating separation cell that forms downstream of the scour lip
626 as the basal high velocity part of the flow is jetted over the
627 depression (Allen, 1971; Farhodi and Smith, 1985; Karim and Ali,
628 2000). This may occur in both subcritical and supercritical flows.
629 When the palaeoflow patterns of element 2 (at K7, K8, and K9) are
630 compared with the streamline patterns of the spindle-shaped
631 erosional marks of Allen (1971), and assuming a scour orientated

632 with the flow direction of the underlying deposits (336°), there is a
633 close fit in terms of variance and spread (Fig. 12). Therefore, the
634 observed palaeoflow patterns can be explained by the presence of a
635 flow separation cell and the generation of reversed bedload
636 transport at the bottom of the flow when passing through the
637 depression (Fig. 12).

638 The second erosional feature located 800 m downstream of the
639 Kleine Riet Fontein scour is similar in architecture and fill. It shows
640 erosion and downcutting surfaces both perpendicular (10° cut to
641 330° palaeoflow) and parallel (335° cut to 330° palaeoflow) to
642 regional palaeoflow. The morphology suggests this second erosional
643 feature is also a large composite scour-fill (>300 m wide). As this
644 northern scour is at the same stratigraphic level, it indicates there
645 may be a larger area of erosional bedforms present. This fits with
646 the interpretations of Jobe et al. (2012) defining the Kleine Riet
647 Fontein area as an area receiving unconfined flows. A spatial
648 distribution of multiple scour-fills in this proximal off-axis area
649 adjacent to distributive channels in Ongeluk River (Fig. 3) supports
650 the interpretation of a channel-lobe transition zone close to the
651 base-of-slope.

652 **6.2 Flow-scour dynamics**

653 The merging of multiple erosion surfaces at the steep and stepped
654 upstream margin of both scour-fills point to their composite origin,
655 with multiple flows shaping the morphology of the basal surface.

656 The basal successions of both scour-fills are similar. However, bed

657 architecture, stacking patterns, erosion surfaces, and facies of the
658 upper elements (3-7) in the Kleine Riet Fontein featurepoint to a
659 more complicated interaction between flow and seabed relief in a
660 later stage of scour evolution.

661 In the Unit A5 feature, the irregular basal surface suggests that after
662 initial development of the scour, the upstream margin was weakly
663 modified, with little evidence of headward erosion. Minor internal
664 erosion surfaces exist, but generally beds taper towards the
665 upstream margin. The stratigraphic transition from the siltstone- to
666 sandstone-prone deposits points to initial sediment bypass (multiple
667 erosion surfaces and medium-grained sandstone lenses) followed by
668 a period of aggradation (structureless Fa1 and banded sandstones
669 Fa3) as the depression filled. The passive depositional character of
670 these packages and the lack of supercritical bedforms, suggests a
671 subcritical nature for the infill.

672 In the case of the Kleine Riet Fontein scour, the evolution of the
673 scour is assessed from the architecture and stacking patterns of the
674 elements (Fig. 13).. The position of the upstream margin of the
675 scour during deposition of element 1 was downstream of the
676 current position of element 2, being deposited after another phase
677 of erosion, evident from the stepped basal erosional surface (Fig.6).
678 Element 3 shows a similar southward migration after reshaping of
679 the updip margin and only partially filled the depression. Element 4
680 is interpreted to have largely filled the accommodation in the
681 downstream and lateral part of the scour. This was truncated by

682 another erosional event that reshaped the updip margin and
683 removed large parts of element 2 and 3. Element 5 and 6 have
684 slightly erosional bases, but mostly infill available accommodation
685 by stacking in a downstream direction. Element 7 has a more
686 uniform thickness but modified the updip margin (Figs. 6 and 13).
687 The interpreted evolution of the basal surface suggests that the
688 initial scouring phase(s) may not be preserved due to sequential
689 deepening and widening of the scour.. The stacking of the elements
690 and internal erosion surfaces in the Kleine Riet Fontein scour-fill
691 indicate upstream migration (Fig. 13) and lengthening of the original
692 scour surface through headward erosion. Headward erosion, or
693 backward incision, occurs in both supercritical and subcritical flows
694 (e.g., Izumi and Parker, 2000; Hoyal and Sheets, 2009). The
695 sedimentary facies and bed geometries of the elements in the upper
696 package are characterised by stoss-side preserved steep climbing-
697 ripple dominated sandstones that indicate rapid localised fallout
698 from relatively low-concentration turbidity currents. Climbing ripple
699 lamination extends across almost the whole length of the scour-fill
700 until in close proximity of the upstream head of the scour. The
701 shallow (a few degrees) upstream depositional dip observed in a
702 number of the infill elements (3, 4, 5) resemble backset bedding (Fig.
703 9). Backset bedding has been linked to abrupt changes in
704 confinement (Ito et al., 2014), associated with the occurrence of a
705 hydraulic jump (Jopling and Richardson, 1966; Lang and Winsemann,
706 2013; Cartigny et al., 2014; Ito et al. 2014). The backset deposits,
707 and the majority of the scour-fill, are characterised almost

708 exclusively by climbing ripple-lamination (Fa 2). These deposits may
709 be either the product of rapid settling from the downstream parts of
710 hydraulic jumps as the scour migrated headward, or the products of
711 subsequent infill by subcritical currents after being previously cut by
712 flows that underwent a hydraulic jump. The measurements of
713 hydraulic jumps over giant scours in a natural gravity current
714 demonstrate, however, that the hydraulic jump enhances upward
715 fluid movement for large distances downstream of the hydraulic
716 jump (Sumner et al., 2013), and therefore minimises sedimentation
717 within the scour. Additionally, the presence of climbing ripple
718 lamination very close to the head of the scour suggests that this
719 final phase was the product of depositional subcritical flows. In
720 combination with the lack of evidence for any depositional features
721 typical of supercritical conditions (supercritical bedforms), other
722 than the backset bedding in elements 3-5, this suggests that the
723 predominant fill of the scour was by subcritical flows. Thus the
724 inception, deepening, and sediment bypass phases of these giant
725 scours may well have been associated with hydraulic jumps in
726 supercritical flows, whilst their infill was dominantly the product of
727 later subcritical flows. This contrasts to the supercritical deposits
728 interpreted in other examples of backset bedding (Jopling and
729 Richardson, 1966; Lang and Winsemann, 2013; Cartigny et al., 2014;
730 Ito et al., 2014).

731 The infill character of the scour to the north of the main Kleine Riet
732 Fontein scour is very similar, with a lower siltstone-prone package
733 and an abrupt change into an upper climbing-ripple dominated

734 sandstone package. This suggests both features share a similar
735 depositional history, which could be due to an internal control
736 linked to an updip avulsion or an external control such as a
737 substantial change in turbidity current energy and/or size. Existing
738 local seabed depressions consisting of partially filled scours could
739 have triggered hydraulic jumps and subsequently reshaped their
740 morphology in a similar manner in both scours, followed by lower
741 energy subcritical flows, explaining the similarity in infill character.

742 **6.3 Preservation of giant scour-fills**

743 Within scour fields in modern systems, amalgamation or
744 coalescence of scours is a common phenomenon (e.g., Parker, 1982;
745 Macdonald et al., 2011a; Fildani et al., 2013; Shaw et al., 2013). It
746 has been emphasized that changes in the behaviour of flows as they
747 pass over the erosional relief of small-scale scours leads to the
748 development of larger depressions (e.g., Shaw et al., 2013). This
749 could be due either to the development of flow separation zones,
750 enhancing erosion, or to the triggering of a hydraulic jump (Sumner
751 et al., 2013). A gap remains, however, in scale between scours
752 documented from the modern seabed and megaflutes and scours
753 interpreted from outcrop studies (Fig. 14). Exhumed scours of the
754 scale described here, and filled with turbidites, have not been
755 described in detail previously. However, the scale of these scour-fills
756 coincides with the range known from modern-day scours (Fig. 14) in
757 CLTZs (e.g., Kenyon et al., 1995; Wynn et al., 2002a; Macdonald et
758 al., 2011a). Due to their composite character and upstream

759 migration, these scour bodies are able to reach significant
760 dimensions. This study shows that within sand-rich turbidite
761 systems, such as in the Karoo Basin, scour-fills of dimensions
762 documented from modern system can be expected to be preserved,
763 and care is needed to discriminate them from submarine channel-
764 fills.

765 Possible explanations for the preservation of large-scale composite
766 scour-fills within CLTZs include i) large scale avulsion of the main
767 feeder system prior to channel propagation into the scoured CLTZ
768 area, ii) the presence of scours at the maximum extent of channel
769 propagation into the basin, and/or iii) lateral position during channel
770 progradation. Differences can however be expected in the character
771 of scour fills between the different preservation mechanisms (Fig.
772 15). As CLTZ's are known to be extensive and widespread areas (in
773 modern systems ~500km - >10,000 km²) (Wynn et al., 2002a) and
774 channel systems are limited in cross-sectional dimensions, it implies
775 that not all scour remnants, including larger coalescent ones, will be
776 reworked by a period of channel propagation (Macdonald et al.,
777 2011a) through a lobe environment (Jegou et al., 2008; Macdonald
778 et al., 2011b; Morris et al., 2014). In the case of a channel
779 progradation adjacent to the scour, the infill signal will depend on
780 the nature of the channel (A1 and B1). In the case the channel is
781 purely erosionally confined (A1), it is expected to show a coarsening-
782 and thickening-upward profile due to the sand-rich nature of
783 overbank deposits (Johnson et al., 2001; Grecula et al., 2003; Van der
784 Werff et al., 2003; Brunt et al., 2012). Progradation of a levee-

785 confined (B1) system could however lead to the reverse infill
786 pattern, as more fine-grained (levee) materials would be deposited
787 inside the scour when the channel-levee system matures. In the
788 case of maximum channel progradation (B2), the overall initial
789 sedimentary signal would be coarsening- and thinning upwards as
790 the scour would have been filled by axial lobe sands during
791 retrogradation. In the case of avulsion of the main feeder channel
792 before complete fill of the scour, the depositional signal would be
793 characterized by a fining- and thinning profile as fringe materials
794 from adjacent lobes are likely to be deposited within the abandoned
795 scours. As the infill signal of both examples here presented fit with
796 infill signal A (coarsening- and thickening-upward) of Figure 15, their
797 depositional history is most likely related to (one of) the two
798 presented preservation mechanisms (B1 and B2).

799 **7. Conclusions**

800 This study reports the first detailed documentation of exhumed
801 giant (>1000-1500 m long) turbidite-filled scours from ancient deep-
802 marine settings. Palaeogeographically, both scour-fills are
803 constrained to base-of-slope channel-lobe transition zone settings.
804 The scour-fills exhibit composite and downstream asymmetric basal
805 erosion surfaces, with internal erosion surfaces and, in one case,
806 evidence for extensive headward erosion. The sedimentary infills
807 show stepped coarsening- and thickening-upwards trends, with the
808 basal fine-grained deposits being associated with low-energy tails of
809 bypassing turbidity currents and subsequent reworking. Palaeoflow

810 patterns within the Fan3 Kleine Riet Fontein scour indicate
811 complicated deflected flow patterns, which supports interpretation
812 of headward recirculation and downstream flow expansion. The
813 simple infill architecture of the Unit A5 scour suggests a rather
814 straightforward cut-and-fill history. The facies and architecture of
815 the Fan 3 Kleine Riet Fontein scour-fill, however, points to a more
816 dynamic history of interactions between flows and the relief of the
817 scour, resulting in a more complicated architecture with evidence
818 for headward erosion and a series of large internal erosion surfaces.
819 The contrast in depositional architecture of the two scour-fills
820 demonstrates that these erosional bedforms can develop diverse
821 depositional histories, presumably as a function of the interaction
822 between the character of the turbidity currents and evolving
823 geometry of the scour surfaces. The steep updip margins, stepped
824 coarsening- and thickening upward successions of dominantly
825 subcritical flow deposits, and internal palaeocurrent dispersal
826 patterns contrast with laterally and stratigraphically adjacent basin-
827 floor channel-fills. Despite their palaeogeographic setting and
828 evidence for formation by hydraulic jumps, their fills, including
829 backset deposits, do not correspond with deposition from
830 supercritical flows. Documenting the facies and architecture of
831 scour-fills is important for the identification and description of areas
832 dominated by sediment bypass in the rock record, and has
833 consequences for the accurate geological modelling of CLTZs.

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1267 **Figure captions**

1268 Fig. 1. Location map of the Laingsburg and Tanqua depocentres
1269 within the Western Cape (South Africa) and schematic
1270 interpretations of the Fan 3 and Unit A5 fan systems (based on
1271 Sixsmith et al. (2004) and Hodgson et al. (2006)). Imaged from
1272 Google Earth.

1273 Fig. 2. Stratigraphic column of the deep-water deposits from the
1274 Laingsburg depocentre and the Tanqua depocentre, based on Pr lat
1275 et al. (2009) and Flint et al. (2011). The fan systems discussed in this
1276 paper (Fan 3 and Unit A5) are highlighted.

1277 Fig. 3. Detailed maps of case study areas with locations of
1278 sedimentary logs, and outlines of Fan3 (A) and Unit A5 (B). Solid line
1279 in A indicates the main profile illustrated in Fig. 6, while the dotted
1280 lines indicate the additional profiles of Fig. 10A. Image from Google
1281 Earth.

1282 Fig. 4. Main sedimentary facies in the case study areas, with (A)
1283 structureless fine-grained sandstone with floating siltstone clast
1284 (Fa1) – Unit A5; (B) Steepening of angle of climb within climbing
1285 ripple-laminated sandstone bed to toss-side preserved climbing
1286 ripples (Fa2)– Fan 3; (C) Banded sandstone (Fa3) – Unit A5; (D)

1287 Lateral discontinuous thin-bedded siltstones and sandstones with
1288 small-scale erosive marks – UnitA5 (Fa4-2).

1289 Fig. 5. (A) Representative facies photographs from the Fan 3 feature.

1290 (A1) Thin-bedded sandstone and siltstone deposits showing the
1291 difference in character below (Fa4-1) and above (Fa4-2) the basal
1292 erosion surface. (A2) Internal truncation within medium-bedded
1293 banded sandstone. (A3) Truncation surface on top of thin-bedded
1294 fine-grained deposits with structured (rippled) sandstone on top and
1295 mudstone clast conglomerate at the base. (A4) Undulating basal
1296 erosion surface truncating thin-bedded deposits. (A5) Photopanel of
1297 the steep stepped southern margin with the locations of A1 and A2
1298 indicated. (B) Representative facies photographs of Unit A5 feature
1299 indicated in Fig.11C with (B1) Thin-bedded siltstones interbedded
1300 with occasional thin coarse-grained sandstone (indicated by white
1301 arrows). Individual siltstone beds show thicknesses >3cm (Fa4-2).
1302 (B2) Small scale soft-sediment deformed sandstones (Fa6). Folds are
1303 indicated y white lines. (B3) Composite erosion surface with initial
1304 mudstone clast conglomerate and banded sandstones, with laminae
1305 parallel to the erosion surface, on top. (B4) Pinchout of sandstone
1306 bed within siltstone thin-beds of the western margin indicated by
1307 the white line.

1308 Fig. 6. Facies correlation panel of main section (solid line in Fig. 3A)

1309 of the erosional feature at Kleine Riet Fontein in Fan 3 with
1310 palaeocurrents shown, with n = number of measurements, μ = mean
1311 palaeoflow and σ = standard deviation. Solid white lines indicate bed

1312 boundaries. The fill is divided into a lower (LP) and upper package
1313 (UP) and a total of seven infill elements as indicated in the bottom
1314 right cartoon. The boundary between the lower and upper package
1315 is indicated by a light blue dashed line. Facies association 1 (Fa1) has
1316 been subdivided into structureless and banded and/or planar
1317 laminated facies; Facies association 2 (Fa2) has been subdivided into
1318 ripple and planar laminated facies.

1319 Fig. 7. Palaeocurrent distribution within the Kleine Riet Fontein area
1320 (Fan 3) subdivided into underlying, fill and overlying deposits.. The
1321 asterisk shows the more detailed stratigraphic change in palaeoflow
1322 direction as also indicated in the main correlation panel of Fig. 6
1323 (K7).

1324 Fig. 8. (A) Fence diagram showing the 3D architecture of the Kleine
1325 Riet Fontein (Fan3) erosional feature. Palaeoflow of the underlying
1326 thin-bedded deposits is indicated (336°). The infill thins-out both in
1327 the eastward and southward directions. See Fig. 3A for log locations.
1328 (B) Detailed log of the fill (K8) showing the division in infill elements
1329 (IE). IE 3 is pinching-out at K8, but is possibly represented by a thin
1330 by-pass interval separating IE2 and IE4.

1331 Fig. 9. Panoramic view of infill element 3 (C), with (C1) Truncation of
1332 elements in southern direction, and (C2) Abrupt bed pinch-out in
1333 the northern direction. Solid white lines indicate bed boundaries,
1334 solid red line indicates the basal erosional surface, and the dashed
1335 red line an internal truncation surface.

1336 Fig. 10. Panoramic photopanel and division in infill elements,
1337 displaying the internal architecture within the Kleine Riet Fontein
1338 erosional feature (Fan 3). The top figure shows seven discrete
1339 sedimentary packages defined as different infill elements. The boxes
1340 show the locations of Inset A & B and Inset C (Fig. 9). The internal
1341 architecture of these elements shows complicated bed architectures
1342 and stacking patterns with abrupt pinch-out of beds in both
1343 southwards and northwards directions. Element 4 (Inset A) and 5
1344 (Inset B) both show stacked bedsets with depositional dips in an
1345 overall southern (updip) direction.

1346 Fig. 11. (A) Panoramic view of the Unit A5 case study showing the
1347 undulating western (updip) margin. (B) Facies correlation panel of
1348 the Unit A5 feature and W15 sedimentary log of scour-fill showing a
1349 coarsening/thickening upward pattern within the infill and a
1350 fining/thinning upward trend on top of the fill. The overlying A5-A6
1351 mud is used as the datum for the sedimentary logs, but is not
1352 included in this panel. (C) Zoomed-in section of the western margin.
1353 Locations of the Unit A5 facies photos within Fig. 5 are indicated.
1354 See Fig. 6 for additional explanatory information.

1355 Fig. 12. Streamlines based on Allen (1971) and possible linkage to
1356 lateral and stratigraphic variance observed at K7, K8 and K9 with $n =$
1357 number of measurements, $\mu =$ mean palaeoflow and $\sigma =$ standard
1358 deviation. These streamlines account for an idealised megaflute
1359 morphology with an orientation of 336° (based on underlying
1360 deposits). See Fig. 6 for log location and exact stratigraphic intervals.

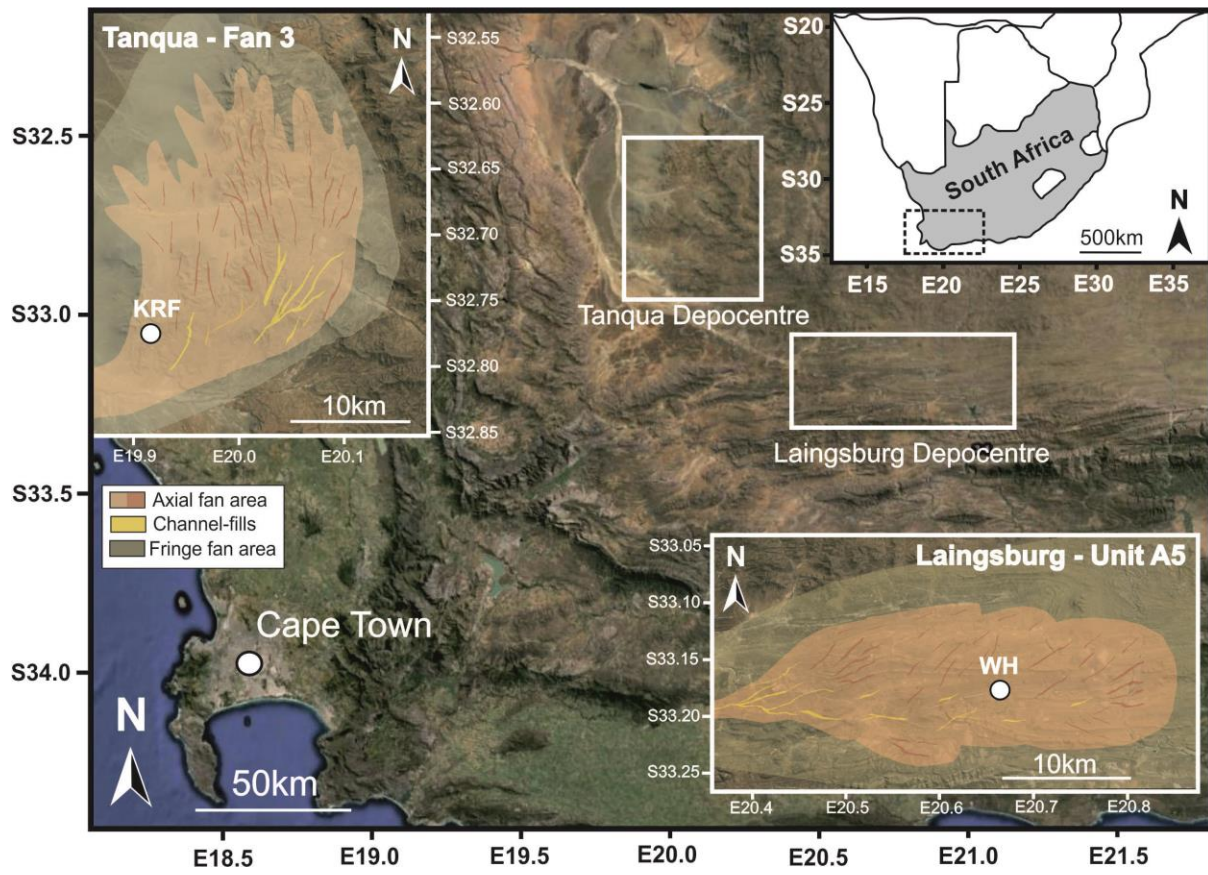
1361 Fig. 13. Depositional history interpretation of infill elements based
1362 on the main profile given in Fig. 6 of the Kleine Riet Fontein (Fan 3)
1363 erosional feature. It is unknown how far the deposits of element 1, 2
1364 and 3 extended on the southern margin as the basal surface of
1365 element 5 has removed a substantial part of this. The palaeocurrent
1366 distribution of the underlying thin-beds is oriented to the NNW
1367 (336°). It must be noted that this section includes a significant
1368 change in orientation around log K6 from 345° (almost parallel to the
1369 underlying palaeoflow) to 030° (transverse to the underlying
1370 palaeoflow), indicated within the map at the bottom right. Imaged
1371 from Google Earth.

1372 Fig. 14. Cross plot of width and depth data of scours and megaflutes
1373 from ancient (outcrop) and modern systems. Scour data from
1374 Macdonald (2011a). Channel trendline is based on Clark and
1375 Pickering (1996).

1376 Fig. 15. Simplified conceptual model to explain alternative ways of
1377 preservation of long-lived composite scours from the initial (T1) to
1378 final depositional setting (T2), divided by vertical infill patterns –
1379 Coarsening and thickening (A) or fining and thinning (B). Two
1380 scenarios are proposed for infill pattern A: A1 – Scour preservation
1381 adjacent to erosionally-confined channel progradation and
1382 successive increase in overbank deposition; A2 – Scour preservation
1383 at the maximum extent of channel progradation followed by lobe
1384 retrogradation. Two alternative scenarios are proposed for infill
1385 pattern B: B1 – Scour preservation adjacent to depositionally-

1386 confined channel progradation and successive development of the
1387 levee; B2 – Scour preservation due to channel avulsion and
1388 successive infill of scours by lobe fringe materials. The '*' indicates
1389 initial position of the middle scour at T1.

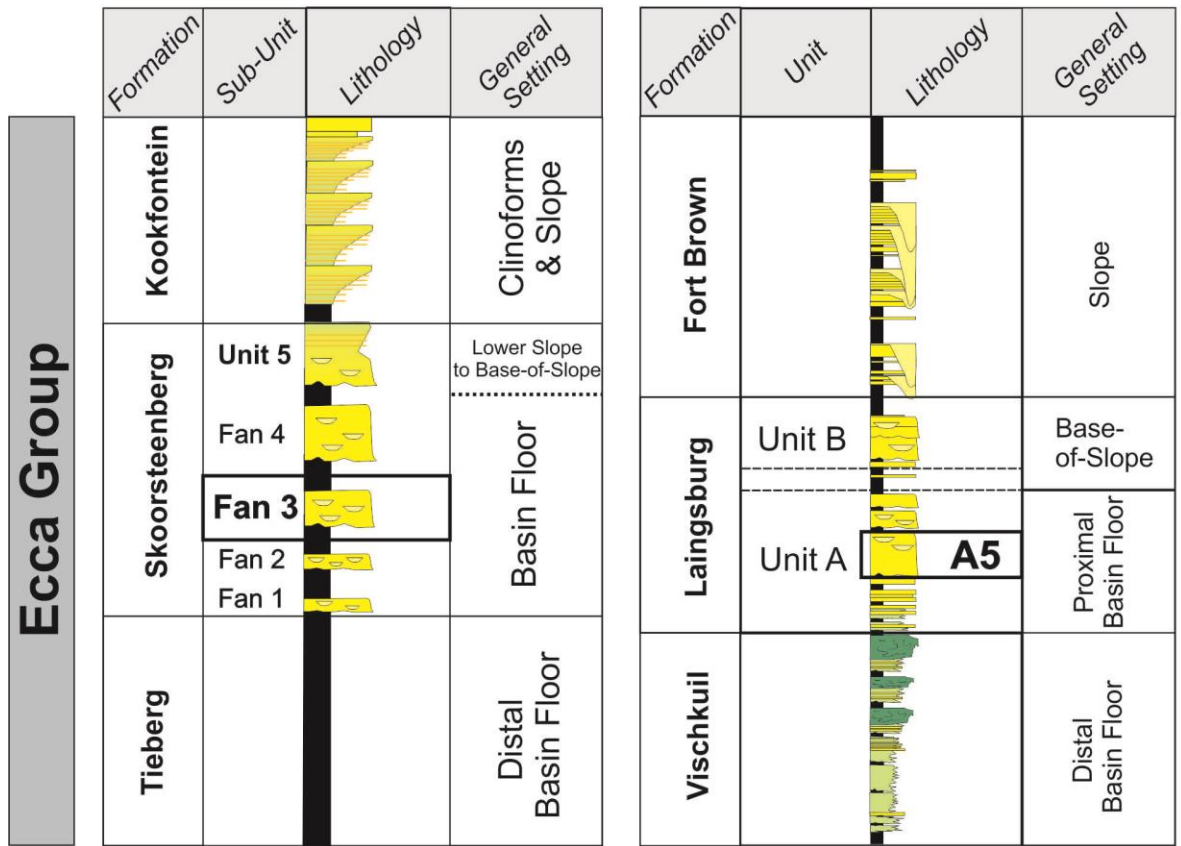
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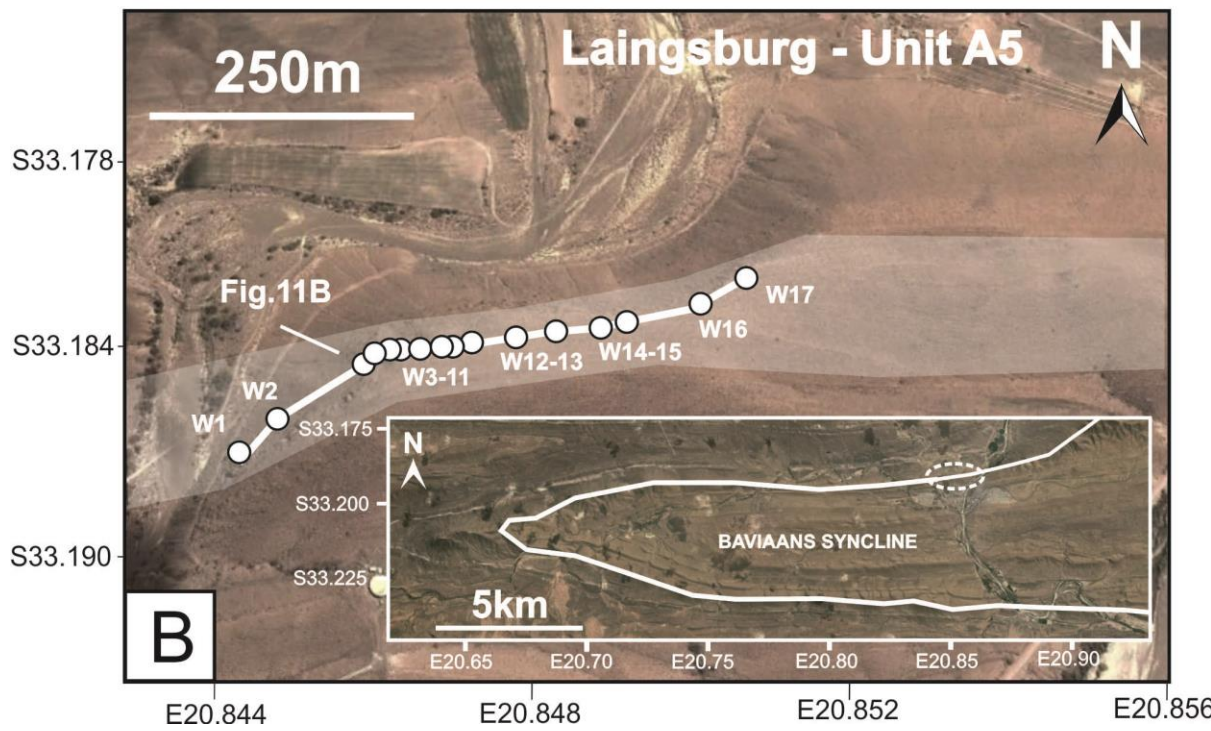
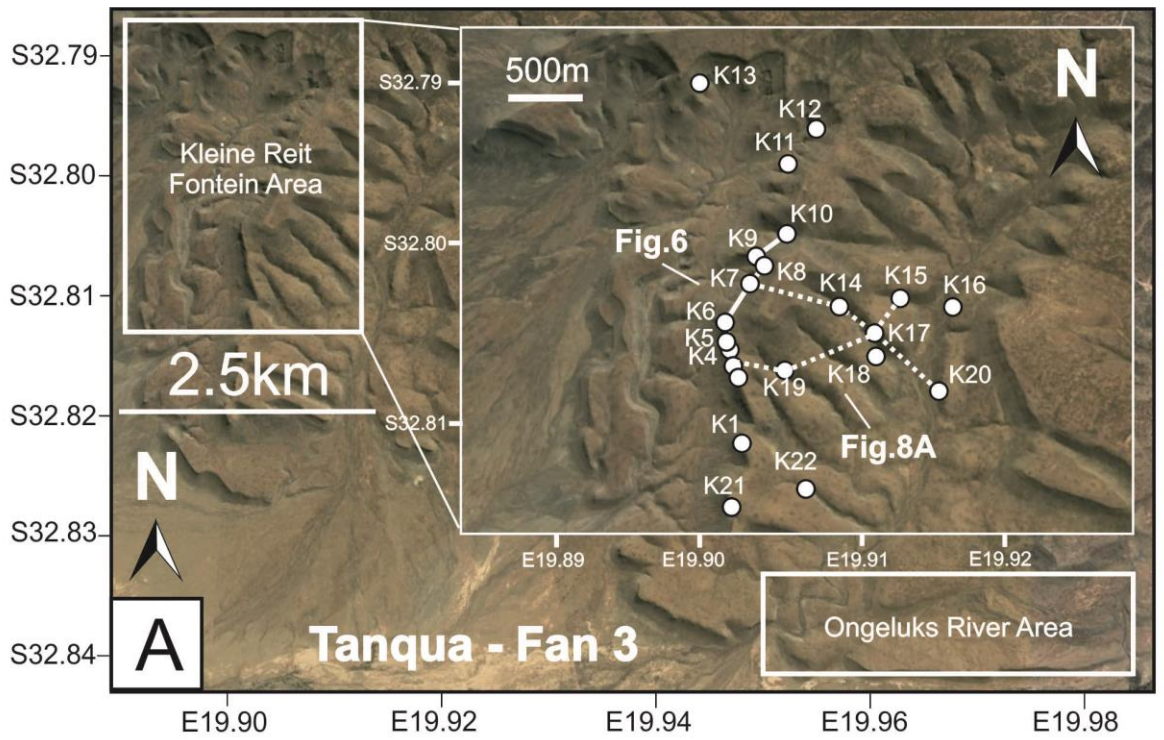
Tanqua Depocentre Laingsburg Depocentre



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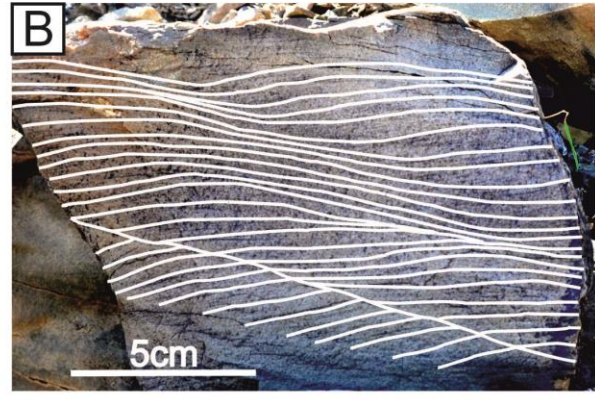
sandstones
 claystones
 siltstones
 mixed silt/sandstones

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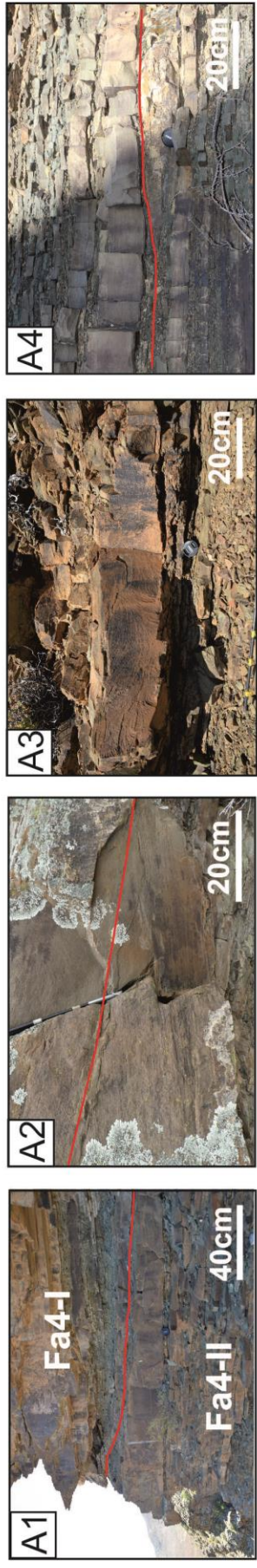
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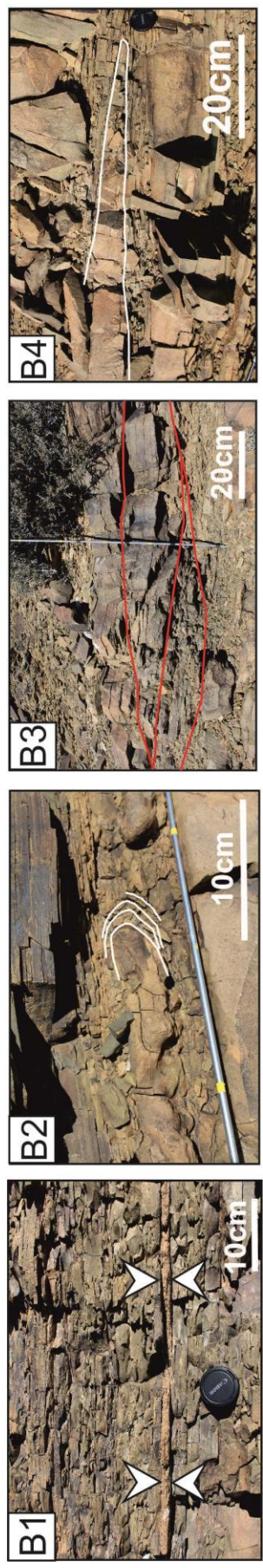
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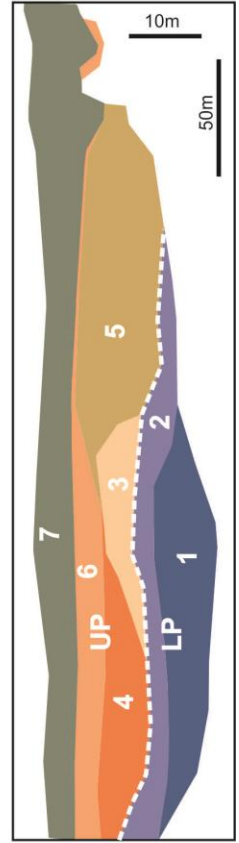
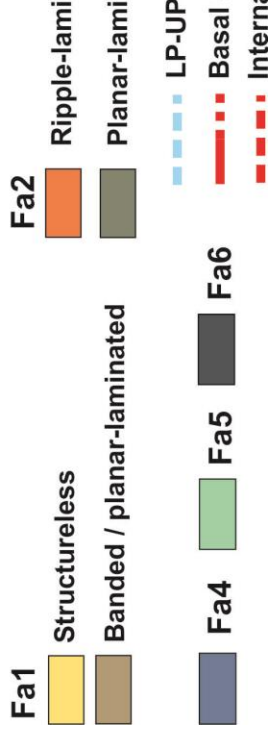
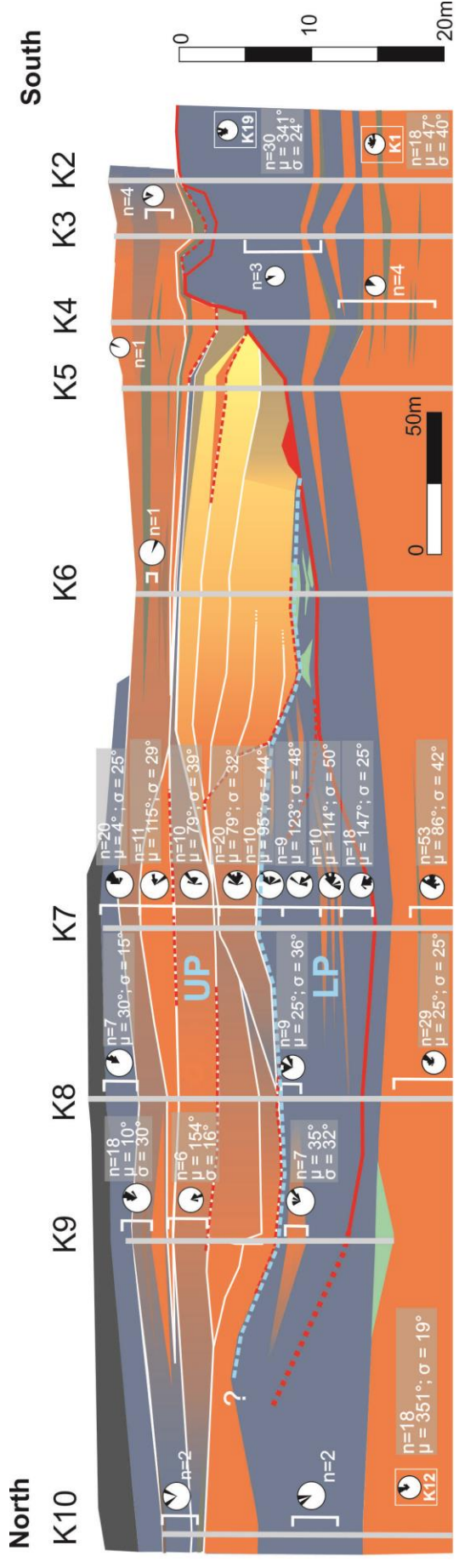
Fan 3



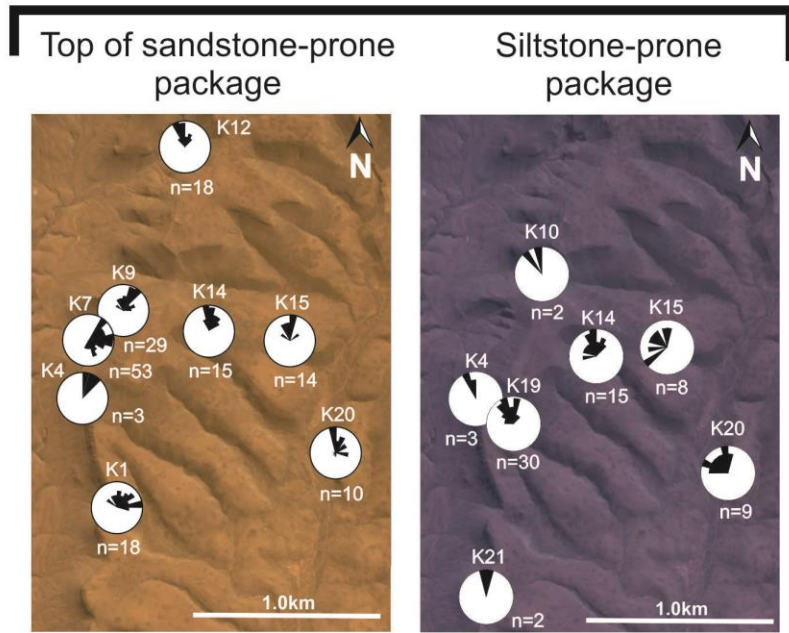
Unit A5

— Erosional surface

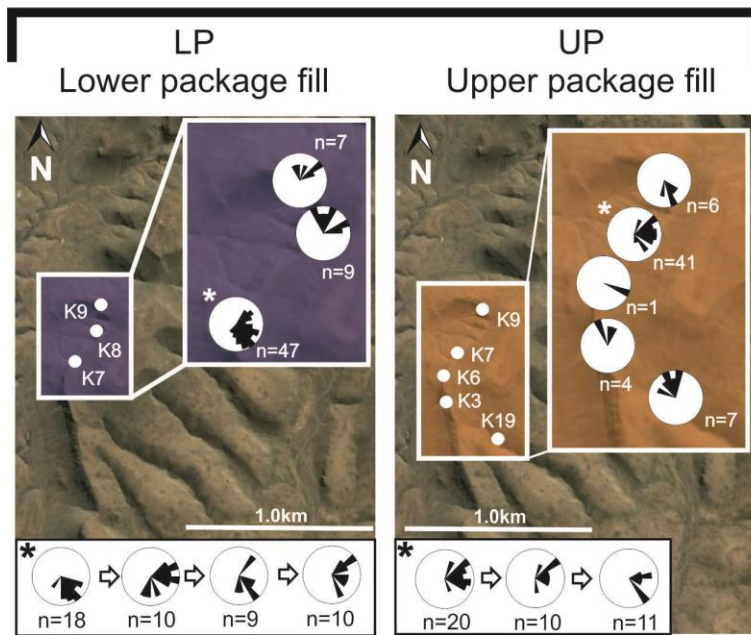




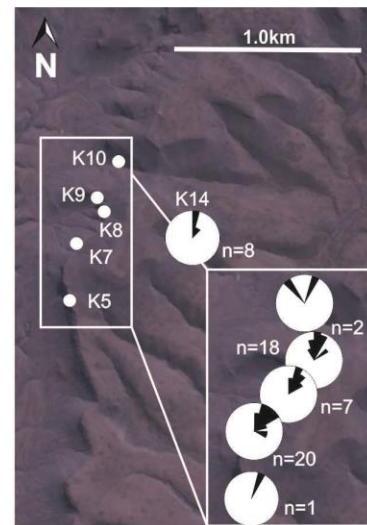
Underlying Deposits



Erosional feature fill

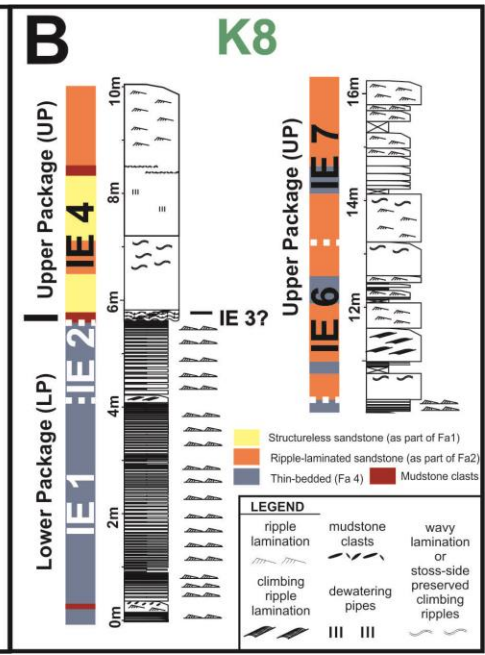
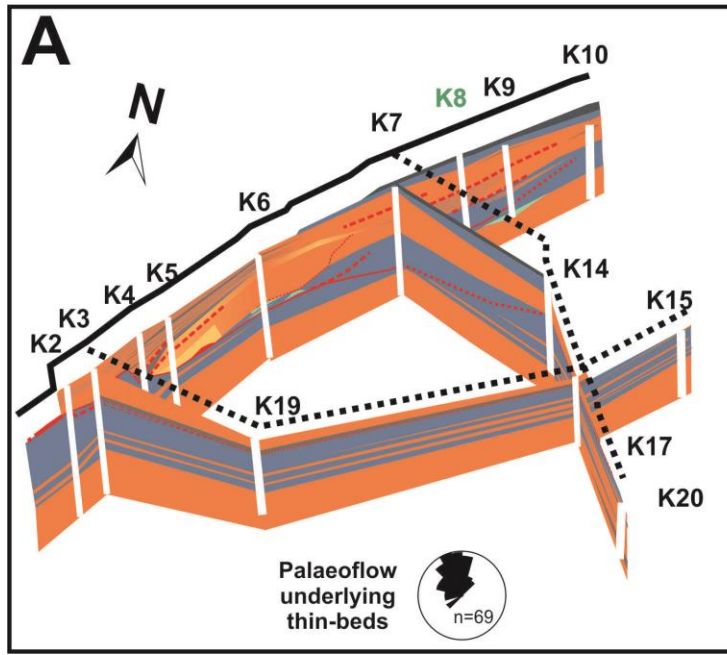


Overlying thin-beds



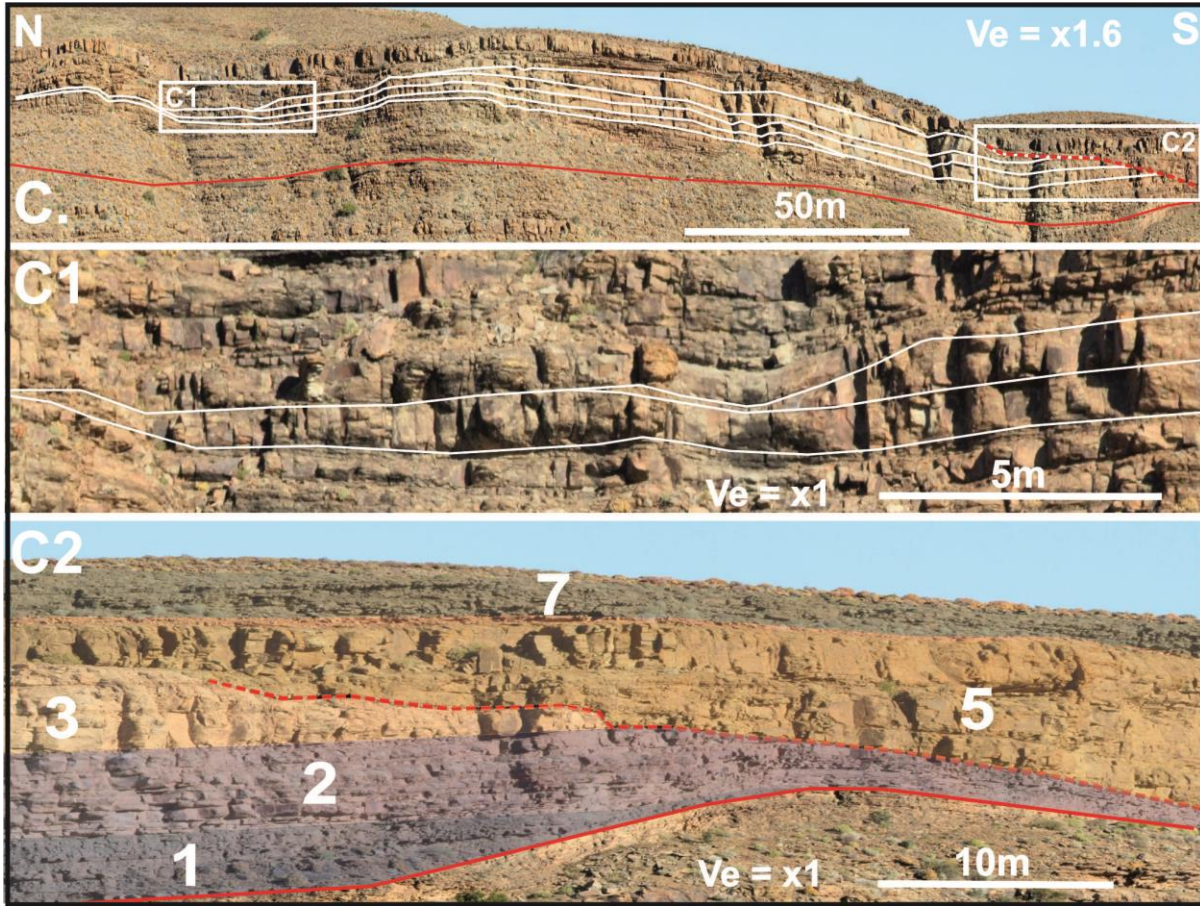
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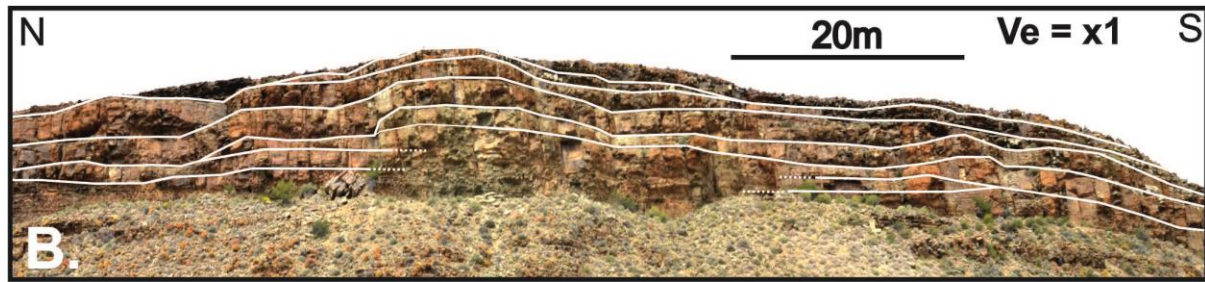
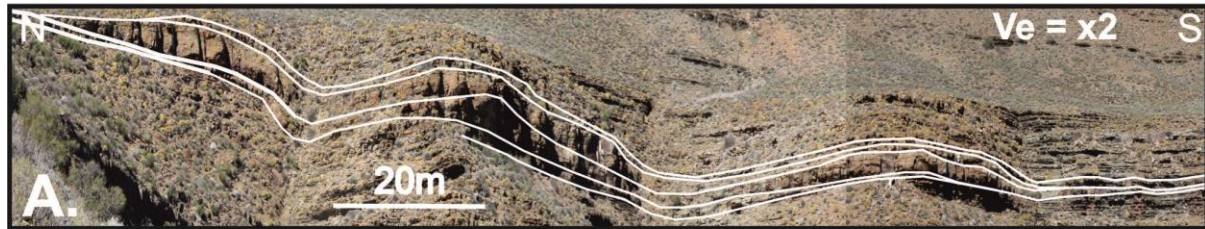
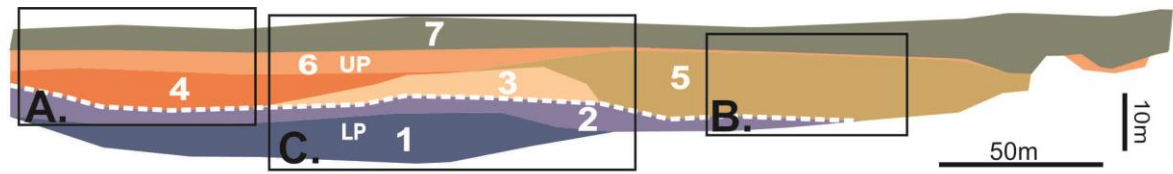
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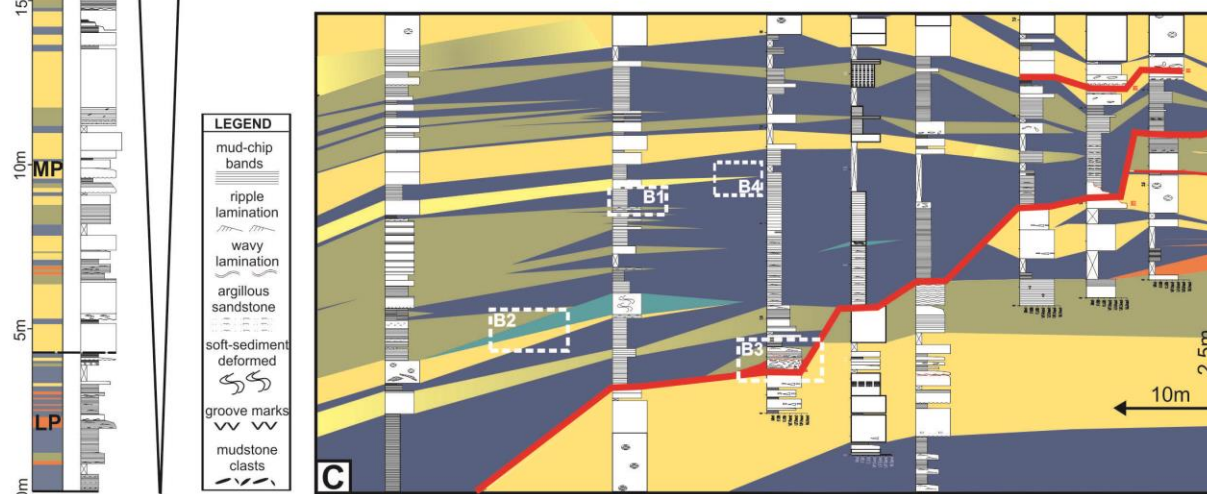
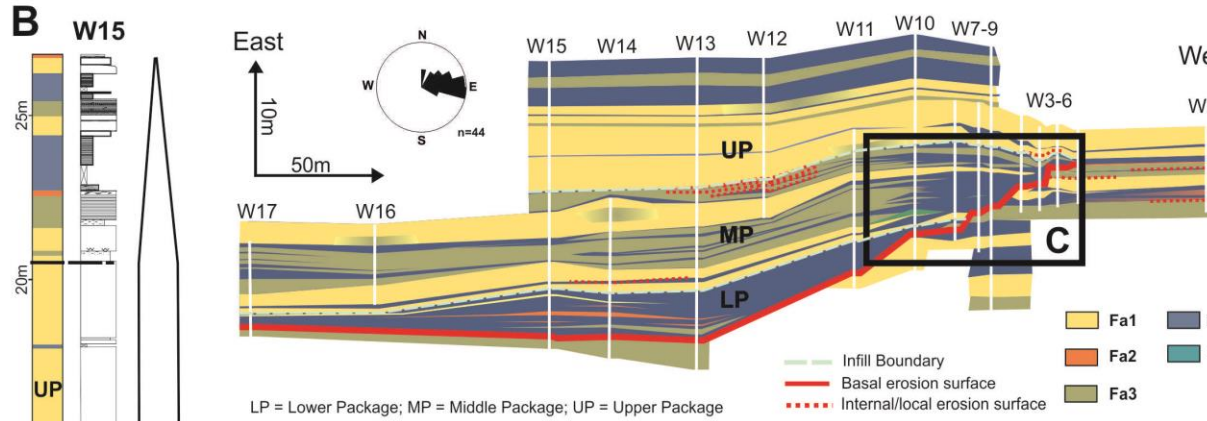
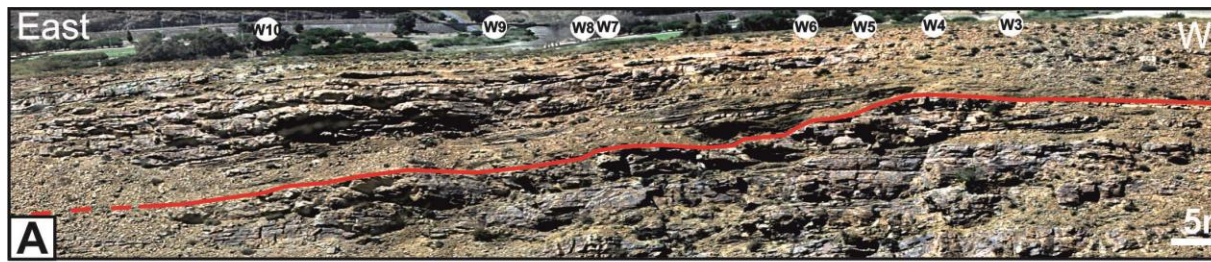
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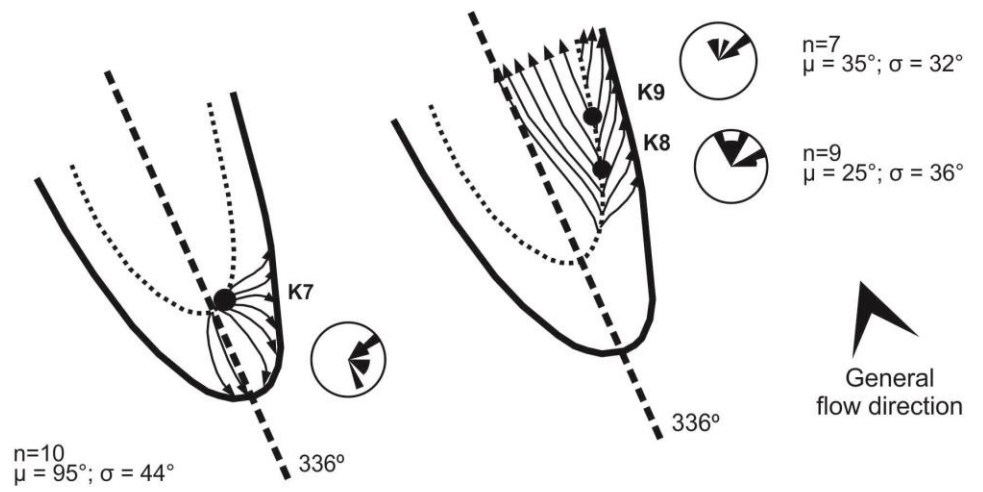
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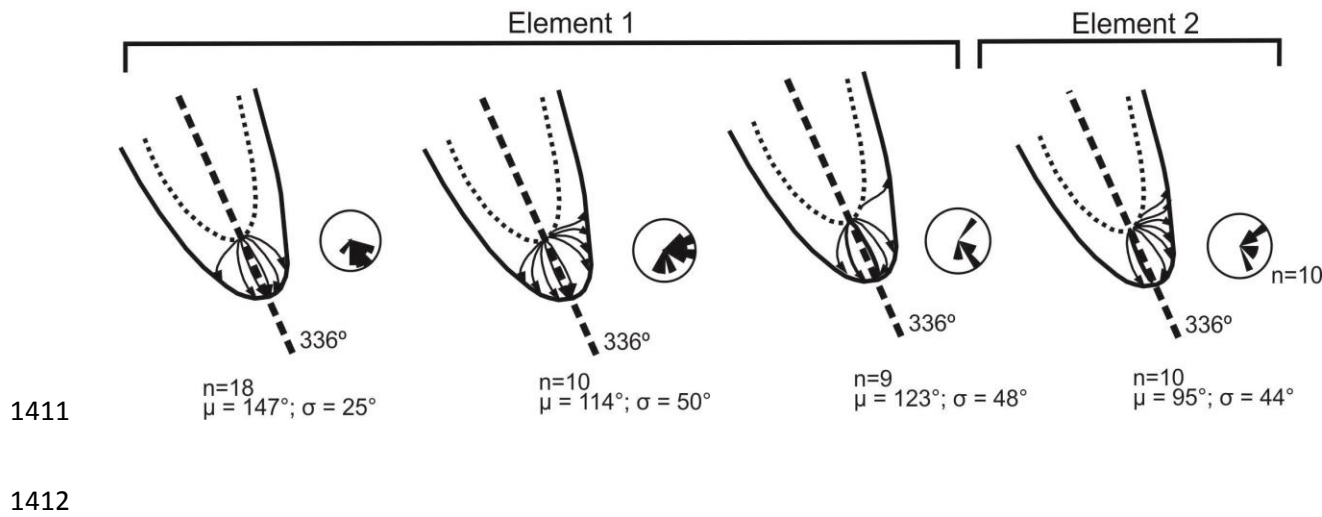
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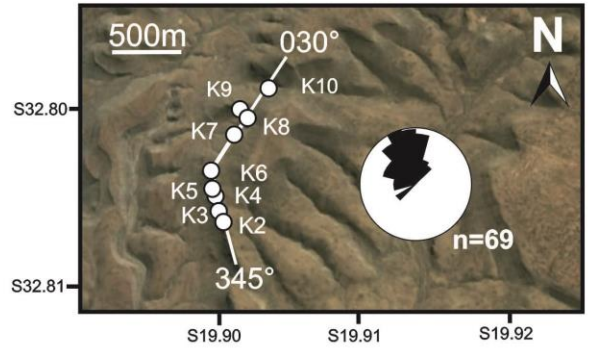
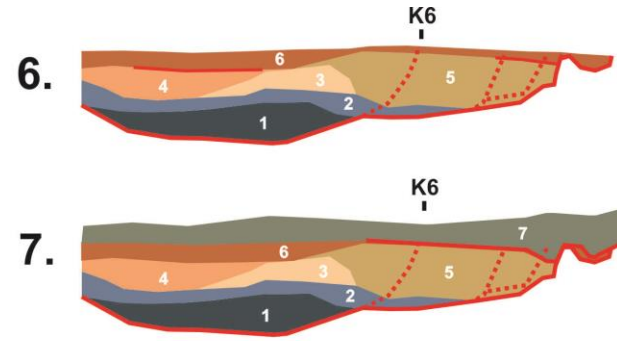
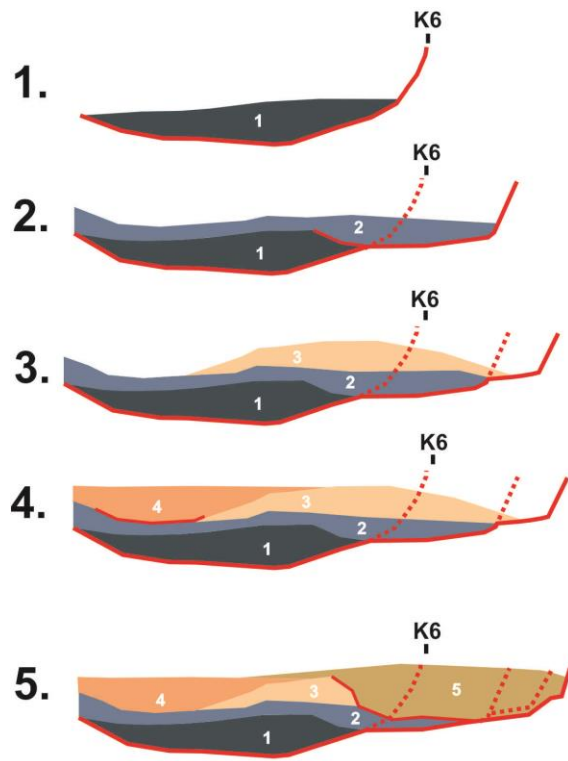
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Lateral - Element 2



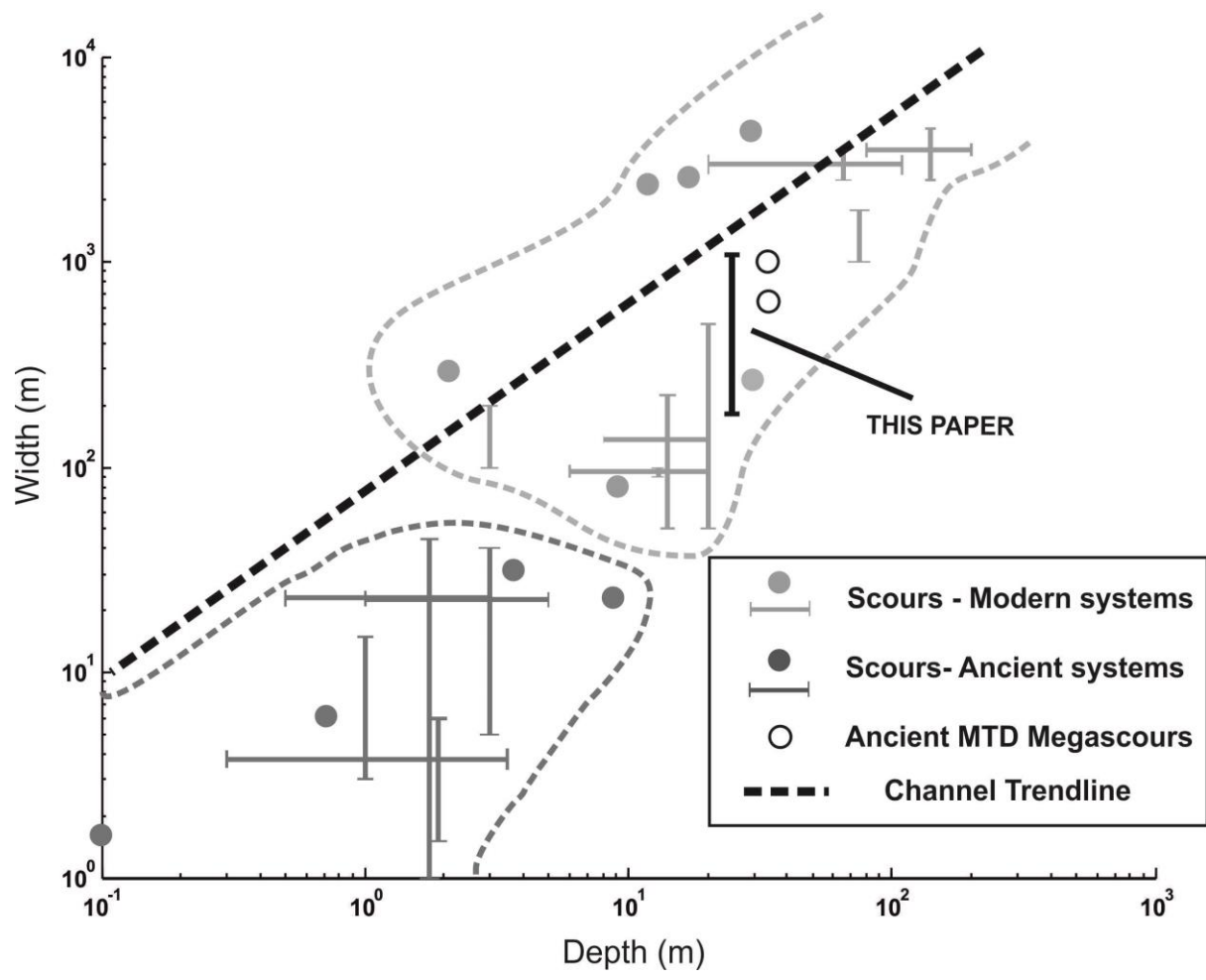
Stratigraphic (K7)





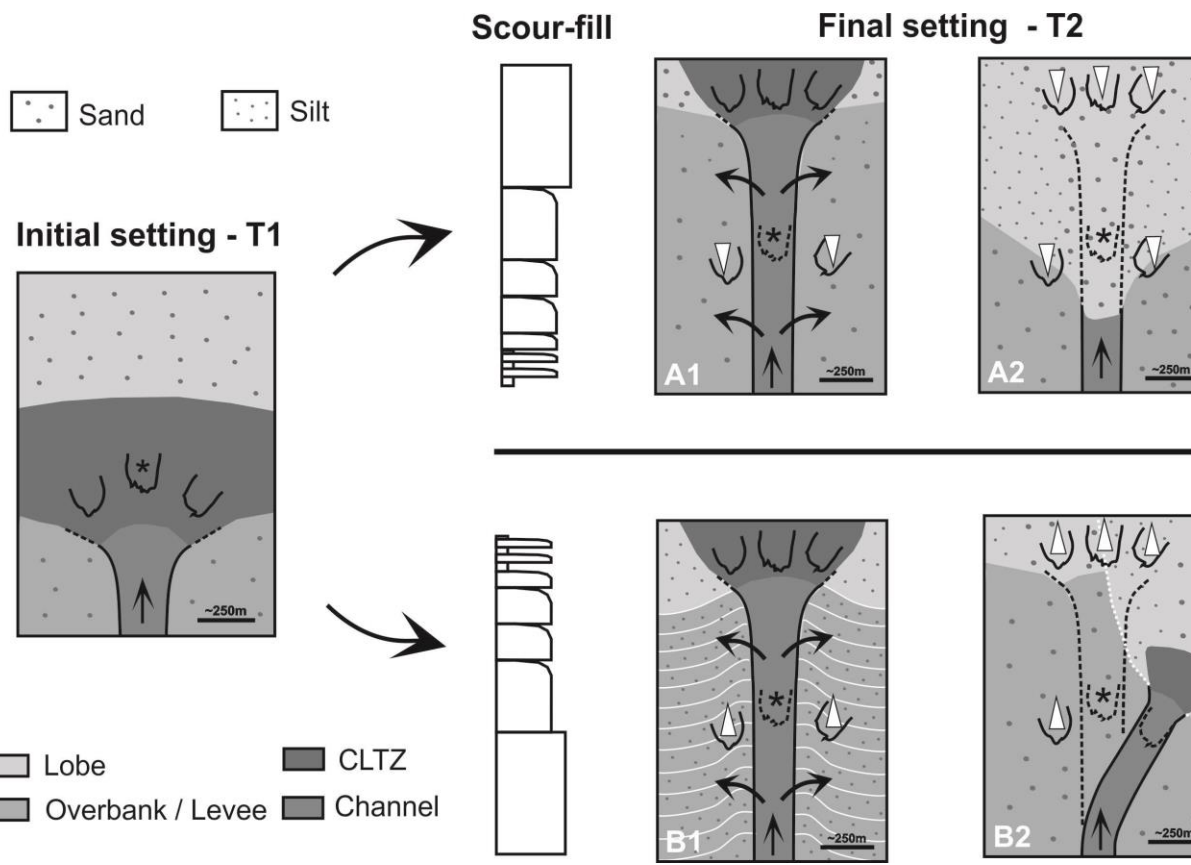
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