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Title: Evolution of a mixed siliciclastic-carbonate deep-marine system on an unstable margin: the Cretaceous of the Eastern Greater Caucasus, Azerbaijan

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19   **ABSTRACT**
20
21   Mixed siliciclastic-carbonate deep-marine systems, herein termed 'mixed systems', are less well
22   documented than their siliciclastic-dominated counterparts, but may be common globally and
23   misinterpreted as transient transition zones between carbonate and siliciclastic deposition. The
24   well-exposed Upper Cretaceous mixed-system of the Buduq Trough, Eastern Greater Caucasus
25   (EGC), Azerbaijan, provides an opportunity to study the interaction between contemporaneous
26   siliciclastic and carbonate deep-marine deposition. The Buduq Trough represents a sub-basin
27   formed within the larger unstable post-rift margin of the EGC. Qualitative and quantitative facies
28   analysis reveals that Upper Cretaceous stratigraphy of the Buduq Trough comprises a
29   Cenomanian-Turonian siliciclastic submarine channel complex, which abruptly transitions into a
30   Coniacian-Maastrichtian mixed-lobe succession. The Cenomanian – Turonian channels are shown
31   to be entrenched in lows on the palaeo-seafloor, with the sequence entirely absent 10 km toward
32   the west, where a Lower Cretaceous submarine landslide complex is suggested to have acted as a
33   topographic barrier to deposition. By the Campanian this topography was largely healed, allowing
34   deposition of the mixed-lobe succession across the Buduq Trough. Evidence for topography
35   remains recorded through opposing palaeocurrents and frequent submarine landslides. The overall
36   sequence is interpreted to represent abrupt Cenomanian-Turonian siliciclastic progradation,
37   followed by ~Coniacian retrogradation, before a more gradual progradation in the Santonian-
38   Maastrichtian. This deep-marine siliciclastic system interfingers with a calcareous system from the
39   Coniacian onwards. These mixed lobe systems are different to siliciclastic-dominated systems in
40   that they contain both siliciclastic and calcareous depositional elements, making classification of
41   distal and proximal difficult using conventional terminology and complicating palaeogeographic
42   interpretations. Modulation and remobilisation also occurs between the two contemporaneous
43   systems, making stacking patterns difficult to decipher. The Buduq Trough is analogous in many
44   ways to offshore The Gambia, NW Africa, and could be a suitable analogue for mixed deep-marine
45   systems globally.

46   **Introduction**

47   *Mixed siliciclastic-carbonate systems*
48
49   Sedimentary successions characterised by contemporaneous deposition of both siliciclastic and
50   carbonate lithologies, herein termed 'mixed-systems', have been identified from the Cambrian
51   (Osleger and Montañez, 1996) to the Quaternary (Dunbar & Dickens, 2003; Tucker, 2003). Mixed
52   systems are formed by a variety of depositional processes (e.g. Mount, 1984; Chiarella et al. 2017)
53   and are consequently recognised in a variety of depositional environments, such as: shoreface
54   (Zonneveld et al. 1997), lagoonal (Mitchell et al. 2001), shelfal (Mount, 1984; Zeller et al. 2015),
55   slope (Gawthorpe, 1986) and(Ditty et al. 1997; Yose & Heller, 1989; Bell et al. 2018; Moscardelli
56   et al. 2019). Mixed-systems deposited in deep-marine (below storm-wave base) are usually formed
57   by material shed from a shallower carbonate-producing shelf that periodically received terrigenous
58   sediment (Fig. 1) (Mount, 1984; Dunbar & Dickens, 2003; Crevello & Schlager, 1980). This
59   material is then re-deposited in deep-marine by a spectrum of sediment gravity flows types, from
60   turbidity currents to submarine landslides (Dorsey & Kidwell, 1999; Miller & Heller, 1994; Tassy
61   et al. 2015; Moscardelli. 2019).

62 *Mixed lobes*

63

64 Sediment-gravity-flows that lose confinement on the slope or basin-floor build lobate depositional
65 bodies, known as lobes, which form important archives of palaeoclimatic and palaeogeographic
66 information (e.g. Hessler & Fildani, 2019). Exhumed deep-marine lobes have been studied in great
67 detail, and a wide variety of stacking patterns, depositional processes and facies distributions have
68 been described and interpreted (e.g. Mutti, 1983; Postma et al. 1993; Prélat et al. 2009; Terlaky et
69 al. 2016; Kane et al. 2017; Bell et al. 2018; Fildani et al. 2018; Fonnesu et al. 2018; Soutter et al.
70 2019, Cumberpatch et al. in prep.). These studies typically focus on siliciclastic systems, with few
71 studies investigating the characteristics of lobes formed in mixed-systems (Fig. 1). This study aims
72 to address this by describing exhumed Cretaceous submarine lobes from the Eastern Greater
73 Caucasus (EGC), Azerbaijan (Fig. 2) which were built by contemporaneous deposition of
74 calcareous and siliciclastic sediment gravity flows. The characteristics of these mixed lobes and the
75 processes that govern their deposition are then compared with siliciclastic lobes. This study will
76 also describe the sedimentological evolution of the basin throughout the Cretaceous, providing
77 insights into the stratigraphic evolution of a basin characterised by unstable margins.

Geological Setting and Basin Structure

Evolution of the Eastern Greater Caucasus

79
80

81 The Eastern Greater Caucasus forms the easternmost extent of the NW-SE trending Greater
82 Caucasus orogenic belt, which runs from the Black Sea in the west to the Caspian Sea in the east
83 (Fig. 2) (e.g. Bochud 2011). The EGC sits on the southern-edge of the Scythian Platform, which
84 represents the southern margin of the Eastern European continent (Saintot et al. 2006). The
85 exposed EGC is mainly composed of Mesozoic-aged sediments that accumulated during multiple
86 phases of extension and convergence related to sequential closure of the Tethys toward the south
87 (Golonka, 2004; Vincent et al. 2007). Most of these Mesozoic tectonic events occurred in the
88 Jurassic, with Upper Triassic-Lower Jurassic compression followed by Lower- to Mid-Jurassic
89 rifting and compression, and Upper Jurassic rifting and compression (Bochud 2011). These
90 tectonic events are recorded by major thickness variations across the Middle Jurassic interval (Fig.
91 3).

92

93 The Lower Cretaceous of the EGC was deposited within an unstable marine environment, as
94 recorded by frequent mass-wasting events and major thickness changes across the interval (Egan
95 et al. 2009; Bochud 2011). Subsidence increased through the Lower Cretaceous and into the early
96 Upper Cretaceous due to back-arc extension associated with the opening of the West Black Sea
97 Basin to the west (Nikishin et al. 2001), resulting in deep-marine deposition of extensive
98 mudstones interspersed by submarine landslide deposits and terrigenous sediments (e.g. Brunet et
99 al. 2003).

100

101 The remainder of the Cretaceous sequence was deposited during a period of thermal subsidence
102 on a southward-dipping slope, with the basin divided into a series of sub-basins (Bochud 2011).
103 One of these sub-basins, the Buduq Trough, encompasses our study area. The Cretaceous
104 stratigraphy is dominated by calcareous and siliciclastic turbidites and conglomerates interbedded

105 with hemipelagic marls and mudstones (e.g. Brunet et al. 2003). A number of intra-Cretaceous
106 unconformities are seen within the basin and are related to periods of compression (Egan et al.
107 2009) or sea-level fluctuations. The end of the Cretaceous sequence is represented by a Base-
108 Cenozoic unconformity formed during Paleogene compression (Bochud 2011).
109
110 Collision of the Arabian and Eurasian plates in the Oligocene (Vincent et al. 2007) deformed the
111 Mesozoic and early Cenozoic succession into a series of exhumed synclines bound by major faults.
112 These faults separate distinct structural zones within the EGC (Fig. 2, Fig. 3) (Bochud 2011).

The Buduq Trough

115 The Buduq Trough is preserved in the east-west trending Qonaqkend structural zone (Fig. 4) and
116 has been interpreted as an Upper Cretaceous 'paleo-valley' incised into Lower Cretaceous deep-
117 marine sediments and Upper Jurassic limestones following a period of compression (Fig. 2B)
118 (Egen et al. 2009; Bochud 2011). It is likely that this compression was related to far-field tectonism
119 in the eastern Black Sea (Sosson et al. 2015), which overprinted the subsidence that characterised
120 the Cretaceous of the EGC. The earliest fill of the Buduq Trough is preserved in the east and is
121 represented by Cenomanian - Turonian sandstones and conglomerates (Fig. 2) (Bochud et al,
122 2011). The nature of this transition varies across and within the Trough; with the Cenomanian-
123 Turonian conformable with the Aptian-Albian at Mt. Kelevudag (Kopaevich et al. 2015) and sitting
124 directly on Barremian at Khirt (Fig. 2, Fig. 3). The overlying Coniacian-Maastrichtian is
125 represented by mixed siliciclastic-carbonate turbidites and is conformable with the Cenomanian-
126 Turonian in the west. In the east, near Cek, the Cenomanian-Turonian is absent, with the
127 Campanian directly overlying Aptian-Albian thin-bedded mudstones, submarine landslide deposits
128 and predominantly siliciclastic turbidites (Fig. 2, Fig. 3). Upper Cretaceous oceanic red beds
129 (CORBs) are also seen throughout the Upper Cretaceous sequence, particularly in the Coniacian
130 – Campanian turbidites and marls, indicating periodically oxic deep-marine conditions (e.g. Hu et
131 al. 2005).

Data and Methods

133 The data set comprises 23 sedimentary logs, totalling 500 m, collected across the Buduq Trough
134 (see supplementary material). Logs were generally collected at 1:25 scale. Bedding and structural
135 data (Fig. 4) and palaeocurrent data (Fig. 5) were collected to ground truth the geological map and
136 cross sections of Bochud (2011). Palaeocurrent readings were quite rare and were taken only where
137 sedimentary structures were clear enough to permit unambiguous data collection. Sparse
138 biostratigraphic data (Bochud 2011) hinders precise correlation across the study area. Chrono-
139 stratigraphic subdivision of the Buduq Trough are still being refined (cf. Bochud 2011; Bragina &
140 Bragin, 2015; Kopaevich et al. 2015), possibly due to the litho-stratigraphic similarities between
141 the units and the complex paleo-topography in which they were deposited (Egan et al. 2009).
142 Therefore we use mapped stratigraphic units (J_1, J_2, K_1, K_2 etc.) and lithostratigraphy to suggest
143 associated ages (Bochud 2011). Sedimentary logs were used to develop a lithofacies scheme (Fig.
144 6, Table 1) and facies associations (Fig. 7).
145 Over 10,000 sedimentological measurements (e.g. bed thickness, grain size, facies) were collected
146 and quantitatively analysed (see supplementary material). Stratigraphic logs were assigned one of

147 seven facies associations (Fig. 7) in order to quantitively compare bed statistics across deep-marine
148 sub-environments (Fig. 8, 9, 10, 11).

Results

Lithofacies

151 Carbonate and siliciclastic lithofacies presented in Table 1 and Fig. 6 represent beds deposited by
152 individual events (event beds) and are classified based on outcrop observations. 'Mud' is used here
153 as a general term, for mixtures of clay, silt and organic fragments.

Facies Associations

156 Facies associations have been interpreted based on the dominant lithofacies (Fig. 6, Table
157 1) and architecture of a given succession and are subdivided into siliciclastic and mixed (carbonate
158 and siliciclastic) associations (Fig. 7). Facies associations FA1, FA2 and FA3 are Cenomanian-
159 Turonian and FA 4, FA 5, FA 6 and FA 7 are Coniacian-Maastrichitan (Bochud 2011). Facies
160 associations commonly used for lobes (Prélat et al. 2009; Spychala et al. 2017) and channels (Kane
161 & Hodgson, 2011; Hubbard et al. 2014) best fit our observations.

Siliciclastic Facies Associations

163 *FA 1: Lobe Fringe*
164 *Observations:* FA 1 is dominated by metre-scale packages of thin-bedded siliciclastic siltstones to
165 fine-grained sandstones with subordinate mudstones and medium-bedded siliciclastic sandstones
166 (Fig. 7A). Beds are laterally extensive for 100's of metres and are commonly flat based and flat
167 topped, often showing normal-grading from fine sandstone to siltstone. Planar and convolute
168 laminations are observed in the upper part of many beds. Debrites, hybrid beds, conglomerates
169 and thick-bedded sandstones are absent.

171 *Interpretations:* Thin-bedded, structured sandstones are interpreted to be deposited from low-
172 concentration turbidity currents (Mutti et al. 1992; Jobe et al. 2012; Talling et al. 2012). The lack
173 of hybrid beds and the thin-bedded nature, lateral-extent, fine-grain size and lack of ripple-
174 stratification indicate deposition in a distal lobe fringe (Fig. 9) (Mutti 1977; Prélat et al. 2009; Marini
175 et al. 2015; Spychala et al, 2017).

176 *FA 2: Channel Axis*
177 *Observations:* FA 2 is composed of metre-scale thick-bedded medium-pebbly sandstones and
178 conglomerates with lesser medium-bedded sandstones and rare thin-bedded sandstones,
179 mudstones, debrites and hybrid beds (Fig. 7B). Within the Cenomanian-Turonian succession, FA
180 2 has the highest frequency of thick-bedded sandstones, conglomerates and bi-tripartite beds (Fig.
181 9). Conglomerates often grade normally into thick-bedded sandstones, commonly associated with
182 a grain size break, with coarse-granular sandstone grade often missing. Where conglomerates do
183 not grade into thick-bedded sandstones they are amalgamated or are less commonly separated by
184 thin beds of mudstone. Conglomerates are poorly-sorted, clast-supported and contain sub-angular
185 – sub-rounded clasts of limestone, sandstone and mudstone that often crudely grade from cobbles
186 to pebbles upwards (Fig. 11). Conglomerates also often contain disarticulated shelly fragments.
187 Sandstone and conglomerate bases are almost always erosional.

188 Thick-bedded sandstones are often normally-graded but can be non-graded or inversely-graded.
189 Decimetre scale mud-clasts are common throughout thick-bedded sandstones and low angle cross-
190 stratification is infrequently observed. Thin- to medium-bedded sandstones often have erosional
191 bases and contain convolute, hummock-like and planar laminations and are normally-graded, with
192 rare examples of inverse- or non- grading. These sandstones are either amalgamated or separated
193 by 10 cm thick mudstone layers, and often contain mud-clasts throughout the bed with granules
194 concentrated at the bed base. Sporadic debrites are also seen within FA 2; these have a deformed
195 mudstone matrix and contain clasts of limestone and sandstone. Hybrid beds are amalgamated
196 into 30-50 cm packages, with individual beds commonly consisting of a thin 2-4 cm fine-medium
197 grained sandstone overlain by a clast and shelly fragment rich 8-12 cm muddy very fine sandstone
198 debrite.
199 'Off-axis' successions have fewer thick-bedded sandstones and conglomerates than FA 2, but
200 more than FA 3, and fewer thin-medium, thick-bedded sandstones than FA 3, but more than FA
201 2 (Fig. 9).
202
203 *Interpretations:* The thick-bedded nature, coarse grain size, amalgamation, erosion and entrainment
204 of clasts within the sandstones suggests that the parent flows were highly energetic and capable of
205 eroding and bypassing sediment (Mutti 1992; Stevenson et al. 2015) and are thus these beds are
206 interpreted as high density turbidites (Lowe 1982). The poorly-sorted nature of the conglomerates
207 suggests that they were initially deposited by laminar flows (Sohn 2000), however apparent grading
208 of conglomerates into thick-bedded sandstones could reflect the transition of hyper-concentrated
209 submarine debris flows into highly-concentrated turbulent flows (Mulder and Alexander, 2001)
210 due to entrainment of ambient water (Postma et al. 1988; Kane et al. 2009).
211
212 The transition from conglomerates to medium-very coarse sandstone is associated with a grain
213 size break, often missing the granule fraction, suggesting bypass of flow (Stevenson et al. 2015).
214 The coarse-grain size and basal location of the conglomerates with respect to thick-bedded
215 sandstones suggests these beds could have been deposited as channel-base lags (Hubbard et al.
216 2014). Erosionally-based lenticular sandstones grading from cobble- to fine-sandstones are
217 interpreted to represent submarine channel fill (Jobe et al. 2017; Bell et al. 2018). This facies
218 association is consistent with gravelly-conglomeratic deposits reported elsewhere to represent
219 submarine channel axis deposition (Postma, 1984; Nemec & Steel 1984; Surlyk 1984; Dickie &
220 Hein, 1995; Kane et al. 2009; Li et al. 2018; McArthur et al. 2019; Kneller et al. 2020).
221
222 While typically related to storm deposits (e.g. Hunter & Clifton, 1982), hummock-like cross-
223 lamination have been interpreted in deep marine environments elsewhere as anti-dune
224 stratification (Mulder et al. 2009), bottom current deposits (Basilici et al. 2012) and reworking of
225 an initial deposit by a subsequent flow (Mutti 1992; Tinterri et al. 2017). The channel axis
226 interpretation of FA 2 speculatively suggests anti-dunes formed by supercritical flows are the most
227 probable interpretation of these hummock-like structures (Araya & Masuda, 2001; Alexander,
228 2008).

229 *FA 3: Channel Margin*
230 *Observations:* FA 3 comprises thin-medium bedded fine-granular sandstones in 30-80 cm packages
231 interbedded with 10-90 cm dark mudstones (Fig. 7C). Within the siliciclastic Cenomanian-

Turonian succession FA 3 has the highest frequency of thin-medium bedded sandstones (Fig. 9). Conglomerates and thick-bedded sandstones are rare in FA 3 (Fig. 9). Thin-bedded sandstones and the upper part of medium-bedded sandstones can be argillaceous, with visible micaceous grains and are often planar, ripple and convolute laminated, with rarer hummock-like laminations. Sandstones are often normally-graded but inverse-grading is also observed. Beds of medium thickness are rich in mud-clasts and commonly amalgamated along mud-clast laden surfaces, bases can be highly erosive and scour-like, removing a significant proportion of the underlying bed. Thin-bedded sandstones can be flat or erosively-based, commonly scoured; where bases are erosional the lowermost part of the bed is commonly rich in granule-grade material (Fig. 7C). Granules and coarser fragments are composed of limestone and sandstone. Infrequent hybrid beds are composed of medium-coarse grained siliciclastic sandstone, overlain by a muddy, occasionally marly fine sandstone debrite.

Interpretations: The thin-bedded nature and presence of tractional structures indicate that this facies association was deposited by a low-density turbidity current (Lowe 1982). Presence of hummock-like laminations could indicate storm-wave influenced deposition (Harms et al. 1975), however their presence within a succession containing thick, dark mudstones and frequent sediment gravity flows suggests a deep-marine origin. Anti-dune formation (Mulder et al. 2009) and tractional reworking of an aggrading deposit (Mutti 1992; Tinterri et al. 2017; Bell et al. 2018) have both been interpreted to form similar hummock-like lamination in deep marine environments. Clean sandstones which grade into argillaceous, micaceous sandstones could indicate transitional flow deposits (Sylvester & Lowe 2004; Baas et al. 2009; Kane & Pontén 2012). The thin-bedded, coarse grain size and erosive nature of these deposits, along with the presence of supercritical bedforms, is similar to the overbank deposits seen adjacent to bypass-dominated channels (Kane & Hodgson 2011; Hubbard et al. 2014; Jobe et al. 2017, Lin et al. 2018, McArthur et al. 2019). These similarities, coupled with the along strike location of FA 3 adjacent to FA 2 (channel axis), has led to the interpretation of FA 3 as a channel overbank (Fig. 9). The lateral transition of FA 2 and 3 is indicative of 'on-axis' to 'off-axis' channel-belt facies (Kane et al. 2009).

FA 4: Lobe Axis

Observations: FA 4 is dominated by > 1 m thick packages of amalgamated conglomerates (Fig. 7D, Fig. 9) interbedded with thin-thick bedded very fine - very coarse sandstones. Within the Coniacian-Maastrichtian succession, the thickest conglomerates are found within FA4 (Fig. 8). The conglomerates are laterally discontinuous, erosionally-based, and are either flat-topped when onlapping, or convex-up when downlapping, the slope (Fig. 7, Fig. 12). Conglomerates increase in frequency, clast size (up to cobble-grade) and thickness, up stratigraphy (Fig. 8) and contain sub-angular to rounded clasts of limestone, sandstone and mudstone (Fig. 11). Within the Coniacian-Maastrichtian stratigraphy the greatest number of amalgamated beds is in FA4 (Fig. 10) and the largest grain size range (majority of beds between very-fine sandstone to medium grained sandstone) is observed (Fig. 8). Within FA 4, a coarser grain size class (of coarse grained sandstone or above) is observed which is almost absent in other Coniacian-Maastrichtian facies associations (FA 5, FA 6, FA 7) (Fig. 8).

275 *Interpretations:* Amalgamation of event beds suggests parent flows were energetic and capable of
276 eroding sediment into the flow (Lowe 1982; Stevenson et al. 2015) and amalgamation of
277 conglomerates indicates deposition in a debris-flow dominated environment (Surlyk 1984; Postma
278 1984; Dickie & Hein, 1995), similar to the debris flow dominated lobes described by McHargue et
279 al. (2019). These conglomerates could also represent sediment bypass within lobe axes (e.g. Kane
280 et al. 2009) or channel fill conglomerates (e.g. Knaust et al. 2014), however their thickness, stacking
281 and geometry are most likely to represent deposition in the axis of a debris-flow dominated lobe.

282 ***Mixed Facies Associations***

283 *FA 5: Lobe Off-Axis*
284 *Observations:* FA 5 is represented by erosively-based thin- to medium-bedded, fine-coarse grained
285 siliciclastic sandstones and thin- to medium-bedded fine-grained calcareous siltstones,
286 conglomerates and fine sandstones (Fig. 7E, Fig. 8). Sandstones with siliciclastic bases and
287 calcareous tops are present throughout and are often amalgamated with siliciclastic and calcareous
288 sandstones, forming packages separated by mudstones and silty-mudstones. Calcareous beds are
289 typically flat-based when overlying mudstones, whilst siliciclastic beds are commonly erosive.
290 Calcareous siltstones and sandstones are massive, whilst siliciclastic sandstones show planar,
291 convolute and ripple laminations, but can also be structureless. Debrites are interspersed, often
292 incorporating thin-bedded calcareous siltstones and sandstones. Hybrid beds are rare (Fig. 9).
293
294 *Interpretations:* The presence of both calcareous and siliciclastic sandstones suggests deposition in a
295 mixed system (Fig. 1, Fig. 10) (Al-Mashaikie & Mohammed, 2017; Chiarella et al. 2017; Walker et
296 al. 2019). Structureless medium-bedded calcareous siltstones and sandstones are interpreted to
297 represent deposition from medium density turbidity currents (Kneller & Branney 1995; Talling et
298 al. 2012; Soutter et al. 2019) aggrading quickly enough to prevent tractional sedimentary structure
299 development in their basal divisions (Kneller & Branney 1995; Sumner et al. 2008). This
300 depositional process is complicated within the calcareous medium-bedded deposits, which appear
301 to have aggraded much more slowly than their siliciclastic counterparts, as evidenced by thin-
302 bedded and medium-grained siliciclastic beds being deposited within medium-bedded and fine-
303 grained calcareous beds. The presence of medium-density turbidites, relatively coarse grain size
304 and common amalgamation suggests lobe off axis deposition (Prélat et al. 2009; Spychala et al.
305 2017).

306 *FA 6: Proximal Fringe*
307 *Observations:* Primarily composed of normally-graded, thin-medium bedded calcareous very-fine to
308 fine-grained sandstones and siltstones, with subordinate thin-bedded siliciclastic fine-medium
309 sandstones and mixed siliciclastic and calcareous sandstones (Fig. 7F, Fig. 8, Fig. 9). Calcareous
310 siltstones and sandstones are flat based when overlying mudstones, but are often erosive at
311 amalgamation surfaces (Fig. 10). Siliciclastic sandstones, either isolated or within mixed beds, are
312 frequently < 3 cm thick, with flat to weakly erosive bases (Fig. 6). Debrites are interspersed within
313 FA 6 and often rework the thin-bedded calcareous siltstones and sandstones. Planar laminations
314 are common within the thin-bedded siliciclastic and calcareous sandstones. Less common ripple
315 laminations show multiple and opposing palaeocurrent orientations. Hybrid beds are rare (Fig. 9).
316

Interpretations: The presence of both calcareous and siliciclastic sandstones suggests deposition in a mixed system (Fig. 1, Fig. 10) (Al-Mashaikie & Mohammed, 2017; Chiarella et al. 2017; Walker et al. 2019). Calcareous sandstones are interpreted to represent deposition from low- to medium-density turbidity currents based on their bed thickness, fine grain size and structuration (Fig. 9) (Lowe 1982; Mutti 1992). The thin-bedded siliciclastic sandstones could represent the depositional products of flow transformation from up-dip debris flows (i.e. the up-dip conglomerates) to turbulent flows following the entrainment of ambient water (Potsma 1988; Haughton et al. 2009), which punctuate slowly aggrading calcareous turbidites, interpreted to represent the remnants of dilute flows (Remacha & Fernández, 2003).

The preservation of both structured and structureless sandstones suggests an off-axis location of deposition; similar preservation of both deposit types has been interpreted in the proximal lobe fringe elsewhere (Prélat et al. 2009; Spychala et al. 2017; Soutter et al. 2019). FA 6 is differentiated from FA 5 based on its thinner beds and less frequent erosional events, and is therefore interpreted as being more distal and deposited within the proximal fringe. Hybrids beds are rare throughout the system therefore a distinction between frontal fringe and lateral fringe is difficult to decipher (e.g. Spychala et al. 2017).

FA 7: Distal Fringe
Observations: Dominated by laterally extensive, metre-scale packages of thin-bedded amalgamated calcareous sandstones which are normally-graded from very fine – fine sandstone to siltstone and are interbedded with metre-scale mudstones and silty-mudstones (Fig. 7G, Fig. 8, Fig. 9). Beds are flat-based, flat-topped and frequently contain both parallel and convolute laminations. Medium-bedded calcareous siltstones-fine sandstones are present, and may reflect amalgamated thinner-beds which are difficult to decipher. Debrites, siliciclastic thin-bedded sandstones and hybrid beds are rare (Fig. 9). The smallest grain size range (between siltstone and very-fine sandstone) is observed in FA6 and FA7 (Fig. 8) and amalgamation is infrequent (Fig. 10). More thin beds are seen in FA7 than elsewhere in stratigraphy (Fig. 7C, Fig. 8, Fig. 9).

Interpretations: Thin-bedded, structured sandstones are interpreted to be deposited from low-concentration turbidity currents (Mutti et al. 1992; Jobe et al. 2012; Talling et al. 2012). The presence of medium-bedded calcareous siltstones-fine sandstones and lack of ripple laminations suggest slow aggradation from a turbulent flow (Remacha & Fernández 2003; Bell et al. 2018). Lack of ripple lamination suggests flows did not reach significant velocity to generate ripple laminations (Baas et al. 2016). The infrequency of hybrid beds and siliciclastic beds within this facies association supports deposition within a carbonate dominated environment and the thin-bedded nature, lateral-extent, fine-grain size and lack of ripple-stratification suggests deposition in a distal lobe fringe (Mutti 1977; Prélat et al. 2009; Marini et al. 2015; Spychala et al, 2017).

Discussion

Nature of the Upper Cretaceous Topography

Toward the west of the Qonaqkend Zone, Upper Cretaceous deep-marine sandstones and limestones can be seen to thin towards, and onlap, Upper Jurassic limestones (Fig. 12, Fig. 13, Fig.

359   14). These Upper Jurassic limestones must therefore have formed 100s of metres of relief on the
360   Cretaceous seafloor. The most likely mechanism for the generation of seafloor topography is an
361   allochtonous block (Fig. 13, Fig. 14, Fig. 15). The presence of decametre-scale allochtonous blocks
362   (mega-clasts of Blair & McPherson 1999) and submarine landslide deposits throughout the
363   Cretaceous stratigraphy indicates a highly unstable margin, supporting this view (Fig. 13, Fig. 15).
364   The identification of a basin-scale submarine landslide deposit, which forms the Qizilqaya and
365   Shagdag mountains toward the west, further validates this interpretation (Bochud 2011) (Fig. 15)
366   with the mega-clasts in the west possibly forming part of this deposit (Fig. 14, Fig. 15, Fig. 16).
367   The contact is therefore formed as the Cretaceous stratigraphy infilled the accommodation present
368   on the irregular surface of the deposit. Such relationships have been observed at outcrop (e.g.
369   Burbank et al. 1992; Armitage et al. 2009, Kneller et al. 2018) and in the subsurface (Fig. 17) (e.g.
370   Soutter et al. 2018, Casson et al. 2020). Differential compaction around these rigid blocks will have
371   resulted in steepening of strata adjacent to the block, which may contribute to the gradual
372   steepening identified (Fig. 12, Fig. 14), which has been reported elsewhere (e.g. Burbank et al.
373   1992).
374
375   ***Upper Cretaceous evolution of the Buduq Trough***
376
377   Deep-marine deposition within the Buduq Trough began following a period of compression and
378   folding in the mid-Cretaceous (Fig. 16) (Egan et al. 2009, Bochud 2011). Evidence of this
379   compression is seen within the earliest fill in the Trough, which is preserved toward the east of the
380   Qonaqkend structural zone. This early fill is represented by Cenomanian - Turonian conglomeratic
381   slope channels that either erode into Barremian deep-marine mudstones or sit conformably on
382   thin-bedded Aptian-Albian siliciclastic turbidites. These basal-Cenomanian stratigraphic
383   relationships are suggested to be caused by channels preferentially infilling lows present on
384   seafloor, forming entrenched channel axes that pinch-out laterally against Barremian mudstones
385   (Fig. 16). These lows may have formed during mid-Cretaceous compression and folding (Egan et
386   al. 2009; Sosson et al. 2016) or through submarine slope failure and consequent scour-formation.
387
388   It is possible that poorly preserved thin-bedded Aptian-Albian turbidites represent the distal
389   extents of the Cenomanian slope channels that were either eroded by the channels during
390   progradation or deposited within isolated lows on the Barremian slope. These lows may have
391   formed in response to similar processes to those which entrenched the Cenomanian channels. The
392   abrupt nature of the transition from distal fine-grained turbidite deposition to conglomeratic slope
393   channels may correspond to either tectonic rejuvenation during the Mid-Cretaceous
394   compressional event (Fig. 16) (Egan et al. 2009) and/or an abrupt relative sea-level fall, such as
395   the eustatic sea-level fall seen in the mid-Cenomanian (Miller et al. 2003).
396
397   Evidence for basinal topography is present during deposition of the Cenomanian – Turonian, with
398   the sequence almost entirely absent 10 km to the west at Cek, indicating the presence of a relative
399   high in this location. Submarine landslide thicknesses also increase toward this high in the
400   Barremian, suggesting the high influenced deposition from the Lower Cretaceous until the
401   Turonian. Previous work has shown the presence of a large Lower Cretaceous submarine landslide
402   toward the west (Fig. 16) (Bochud 2011), which is likely to form the high and the complex

stratigraphic relationships described previously (Fig. 13, Fig. 14, Fig. 15). It is also likely that this submarine landslide, and other more minor ones in the area, were emplaced during an earlier period of tectonism and instability related to Lower Cretaceous compression (Fig. 16). Evidence for topography (Fig. 12) in the Late Cretaceous is also evident on a smaller scale through paleocurrent reversals in low-density turbidites (e.g. Kneller et al. 1991) indicating a northward-dipping slope confining southward-directed flows (Fig. 5, Fig. 12), and through the deposition of Upper Jurassic blocks within the Turonian succession, indicating slope instability during this period (Fig. 13, Fig. 15).

Following the Cenomanian-Turonian regression the Trough begins to deepen again during the Coniacian-Maastrichtian (CM), as represented by the deposition of laterally-extensive, thin- to medium-bedded, mixed-siliciclastic-carbonate turbidites overlying the slope channels (Fig. 16). The mixed-lithology of the turbidites contrasts with the dominantly siliciclastic Aptian-Albian turbidites underlying the slope channels, indicating a change in source or paleogeography between the Lower and Upper Cretaceous (Fig. 5, Fig. 16). The presence of thinning and facies changes toward present-day syncline margins, frequent debrites and out-runner blocks, and divergent palaeocurrent distributions indicates that basinal topography had an impact on CM deposition (Fig. 5, Fig. 12, Fig. 13, Fig. 15). This topography may have been formed by differential compaction over the rigid limestone mega-clast, or external compression (Fig. 14, Fig. 15, Fig. 16). Erosional contacts are seen within the CM succession at the base of small, metre-scale channel fills, which occur with increasing frequency through time. These small channel fills are filled by conglomerates and high-density turbidites with similar compositions to the underlying and much more extensive slope channels. The channels are therefore interpreted as small distributary channels in the axes of lobes that formed at the distal ends of the underlying slope channels (e.g. Normark et al. 1979). The increasing frequency and thickness of these conglomerates through the CM (Fig. 8) may therefore represent gradual progradation of the slope channels following their abrupt backstep at the end of the Turonian. Clasts within these younger conglomerates are also more limestone-dominated, which fits with the transition to a more carbonate-dominated system through the Upper Cretaceous (Fig. 11, Fig. 16, Fig. 18).

Mixed-deep-marine deposition continues in the Buduq Trough throughout the remainder of the Cretaceous until Palaeogene compression ceases deposition (Bochud 2011), forming an unconformity between the Upper Cretaceous and overlying Palaeogene and Neogene sediments (Fig. 2, Fig. 3).

A Subsurface Analogue for the Buduq Trough

A seismic-scale equivalent of a mixed-system analogous to the Cretaceous Buduq Trough has been identified and is used to support and increase the resolution of our outcrop-based model. The continental margin offshore The Gambia, NW Africa, developed through the Late Cretaceous with remarkable similarities in timing and evolution to the Buduq Trough (summarised in Casson et al. 2020 in press; Fig. 17). Unconfined mixed-systems developed on the deep-marine basin floor are interpreted to have been line-fed through a heavily canyonised unconformity surface (Fig. 17C). Seismic geomorphology reveals the interfingering of siliciclastic-dominated and carbonate-

445 dominated systems (i.e. at X and Y Fig. 19), similar to that observed on facies and facies
446 architecture scale in the EGC (e.g. Fig. 6, Fig. 7).
447 Sediment gravity flows through the canyons eroded into the underlying carbonate platform
448 redepositing hundreds of metre-scale, seismically-resolvable carbonate mega-clasts 20+ km from
449 the escarpment (Fig. 17B, D); our field work suggests that these blocks may be associated with a
450 multitude of different types and sizes of submarine landslides and blocks that are below seismic
451 scale (Fig. 13, Fig. 15). The presence of carbonate blocks and lobe-architecture in the carbonate-
452 dominated systems (*sensu* McHargue et al. 2019) suggests deposition by debris-flows (i.e. FA 4).
453 Hence two stages of mixing occurs, firstly during erosion to form mixed lithology flows, and then
454 through deposition of interfingering systems. Pervasively channelised siliciclastic-systems with
455 single feeder channels show a distinct seismic geomorphological response to their carbonate
456 counterparts (Fig. 17D, E). The lateral location along the margin of siliciclastic-dominated systems
457 is conceivably related to sediment input points (i.e. shelf-incising canyons) capturing an extra-
458 basinal source of siliciclastic sediment from the shallow marine environment, away from shelfal
459 carbonate factories. Basin floor topography is created by early deposits and influences subsequent
460 lobe deposition (Fig. 17), causing stacking and lateral migration of lobes, which cannot be resolved
461 in the Buduq Trough (Fig. 18) probably because the scale of the study area is smaller than the scale
462 at which migration occurred.
463
464 Documentation of ancient subsurface mixed-systems has been achieved from the interpretation
465 of seismic reflection data (e.g. Moscardelli et al. 2019, Casson et al. 2020). It may also be possible
466 that transitions from calcareous-dominated to siliciclastic-dominated deep-marine systems, which
467 are commonly associated with the rapid arrival (progradation) of the siliciclastic system (e.g. Scott
468 et al. 2010; Kilhams et al. 2012; 2015; Soutter et al. 2019; Cumberpatch et al. in prep.), may have
469 been overlooked as 'transition zones', and in fact represent short-lived mixed systems, which are
470 often below the scale of seismic resolution. The role of mixed-system interactions on a grain-scale
471 and its implications in terms of reservoir quality remain unclear until such systems are drilled
472 (Chiarella et al. 2017; Bell et al. 2018; Moscardelli et al. 2019).

473   ***Mixed lobes***

474 ***Lobe sub-environments:*** If individually observed, the siliciclastic system within the mixed
475 succession could be interpreted as stacked lobes with axial, off-axial and fringe sub-environments
476 identified. The calcareous system, however, would be interpreted as being predominantly lobe
477 fringe deposition (Remacha & Fernández 2003; Bell et al. 2018). Since the two systems are mixed
478 it is difficult to assign a single lobe sub-environment to a sequence of beds as they represent the
479 inter-fingering of two systems (Fig. 19). Due to the interaction of these systems, siliciclastic lobe
480 elements are likely to occur within calcareous lobe elements (Fig. 1, Fig. 17, Fig. 19) (Prélat et al.
481 2009), forming stacks of mixed event beds (D, Table 1). This is further complicated by often highly
482 erosive siliciclastic turbidity currents which can rework calcareous beds, as evidenced by calcareous
483 rip-up clasts within siliciclastic turbidites. This may remove individual calcareous lobe elements
484 from the rock record, and make stacking interpretations more difficult (Fig. 18) (Braga et al. 2001).
485
486 Due to these complexities it is perhaps necessary to refer to such systems with a more specific
487 descriptor (e.g. mixed axis-fringe), or broadly refer to them as 'mixed systems' in order to allude

488   to their complexity and contrast them from siliciclastic-dominated systems (Fig. 19). Use of the
489   siliciclastic lobe hierarchy of Prélat et al. (2009) is possible in mixed systems, but calcareous and
490   siliciclastic descriptors are required (Fig. 19). It is possible to decipher the different systems in our
491   field and subsurface examples, due to their lithological differences being visually resolvable at
492   outcrop (Fig. 6, Fig. 7) and showing different seismic characteristics in the subsurface (Fig. 17, Fig.
493   19). However, without detailed provenance and geochemical analysis it would be very difficult to
494   decipher the mixing of two siliciclastic systems or two calcareous systems, due to similarity in
495   depositional facies and thus seismic character. Unless an individual system can be followed from
496   source to sink in outcrop or the subsurface we must always consider the possibility of multiple
497   systems interacting, modulating each other and complicating stacking patterns (Fig. 19).
498
499   *Stacking patterns:* Deep-marine stacking motifs can show either aggradational, progradational,
500   retrogradational or unorganized stacking patterns (Stow & Mayall, 2000; Deptuck et al. 2007;
501   Straub et al. 2009; Prélat & Hodgson 2013), which can be modulated by both external and internal
502   processes (e.g. Ferguson et al. 2020). Our study shows that in mixed systems it can be difficult to
503   decipher stacking patterns within each individual system due to the convolution of each system by
504   the other (Fig. 18, Fig. 19). Bed thickness trends within the calcareous turbidites are difficult to
505   decipher, possibly due to their narrow grain size range preventing the identification of thinner-
506   beds, and amalgamation within thicker beds (Fig. 18).
507
508   Siliciclastic conglomerates become more frequent and thicker throughout the Coniacian-
509   Maastrichtian, perhaps reflecting a progradation of the siliciclastic system (Fig. 8, Fig. 16).
510   However, bed thickness and grain size analysis for the Coniacian-Maastrichtian do not show any
511   thickness trends or stacking patterns within the calcareous or siliciclastic turbidites (Fig. 18). This
512   suggests that in mixed systems it may therefore not be possible to describe the progradation or
513   retrogradation of an individual system, and only possible to describe the relative ratio between the
514   two; the apparent dominance of the mixed system (e.g. if siliciclastic (s) > carbonate (c) this could
515   be due to progradation of s or by the retrogradation of c, both of which are controlled by a number
516   of external and internal forcings).
517
518   On the scale of the outcrops (100s m), the calcareous turbidites appear to be sheet-like, while the
519   siliciclastic turbidites show thickness variation, representing more typical channel and lobe
520   geometries (e.g. Prélat et al. 2009). Conglomerates observed in the FA4 appear to be confined to
521   isolated depocentres and pinch-out across meters - 10s of meters, indicating the presence of subtle
522   topography (Fig. 12). This suggests the deposition of the conglomerates may have been controlled
523   by depositional topography (compensational stacking) and that the underlying calcareous
524   turbidites do exhibit subtle, long-wavelength thickness changes over a greater scale than observed
525   at outcrop, influencing subsequent sediment routing. Alternatively, the thinning of conglomerates
526   was due to the basinal topography present at this time, preventing these highly-concentrated flows
527   running-out over great distances (Fig. 12).
528
529   *Mixed-system origin:* Previous work on mixed systems has correlated alternations in calcareous
530   and siliciclastic turbidites to 3rd order sea level cycles (Yose & Heller, 1989; Miller & Heller, 1994);
531   the alternations in the Buduq Trough are lower frequency than these cycles but could be
532   interpreted as fifth-sixth order sea level cycles (parasequences) occurring on a 10,000 – 100,000

533 year cycle (Fig. 16) (Van Wagoner et al. 1990), related to Milankovitch orbital cycling
534 (Goldhammer et al. 1990; D'Argenio et al. 1999). Elsewhere mixed systems have been interpreted
535 to represent alternating cool-wet and cold-dry climate cycles driven by precession orbital cycles
536 (García-García et al. 2009). No obvious stacking can be deduced in the study area (Fig. 18)
537 preventing a confident interpretation to be made regarding the forcings behind the high-frequency
538 lithological variations.

540 Rugose carbonate platform margins (e.g. Saller et al. 1989; Grant et al. 2019, Casson et al. 2020),
541 like those observed in the Buduq Trough (Fig. 13, Fig. 16, Fig. 17), have been proposed as conduits
542 for siliciclastics without requiring a sea level change (Francis et al. 2008; Braga et al. 2008; Puga-
543 Bernabéu et al. 2014; Al-Mashaikie & Mohammed, 2017; Walker et al. 2019). This could indicate
544 that the calcareous deep-marine system in the Buduq Trough is part of a much more extensive
545 and line-fed system derived from shedding of active carbonate factories perched on the shelf (e.g.
546 Fig. 17). The contemporaneous siliciclastic system may therefore have been derived from multiple
547 point source conduits along this margin that either 1) periodically punctuated this larger carbonate
548 system or 2) were long-lived conduits permanently bound by carbonate factories (Fig. 16) (Mueller
549 et al. 2017; Moscardelli et al. 2019). Two different sources for separate components of a mixed
550 system have been documented elsewhere (Fig. 1, Fig. 17, Fig. 19A) (Ditty et al. 1997; Riaz Ahmad
551 & Jamil Afzal, 2012; Poprawski et al. 2014; 2016; Chiarella et al. 2017). The presence of Late
552 Jurassic blocks (Fig. 13, Fig. 15) within the Cretaceous complicates this model, with the blocks
553 interpreted as either 1) Late Cretaceous failures from an exposed Jurassic shelf, 2) out-running
554 blocks from Lower Cretaceous failures (e.g. De Blasio et al. 2006) that were subsequently deposited
555 around during the Late Cretaceous, or 3) blocks that were periodically shed through the Late
556 Cretaceous from high-relief Lower Cretaceous slope submarine landslides identified in the west
557 (Fig. 14, Fig. 16).

559 Palaeoflow indicators are limited for the calcareous system due to lack of ripple lamination
560 developed in its fine-grained, slowly accumulating deposits (Baas et al. 2015). It is therefore
561 difficult to decipher whether these siliciclastic and calcareous systems were perpendicular, oblique
562 or parallel to each-other. The palaeoflow indicators that were collected, however, are consistent
563 with a provenance to the north (Fig. 5, Fig. 12, Fig. 15) A northern provenance is also suggested
564 from palaeographic maps for the interval, suggesting a Scythian platform source area (Nikishin et
565 al. 1998), and unpublished provenance data (Vincent, pers.comm.).

Conclusion

568 This study uses the Upper Cretaceous Buduq Trough, Azerbaijan to document the characteristics
569 of an unstable and mixed siliciclastic-carbonate system. Deposition in the Trough is represented
570 by a Cenomanian-Turonian submarine channel complex, which transitions into a Coniacian-
571 Maastrichtian mixed lobe succession. This sequence represents an abrupt Cenomanian regression,
572 probably related to a mid-Cretaceous compressional event and/or an abrupt mid-Cenomanian
573 eustatic sea level fall; followed by a relatively abrupt late Turonian-Coniacian transgression, likely
574 associated with subsidence caused by back-arc extension. Throughout the remainder of the
575 Cretaceous, the mixed-system exhibits weak progradation. A westerly topographic high formed by
576 a Lower Cretaceous submarine landslide complex deposited during earlier compression is

interpreted to have prevented deposition of the Cenomanian–Turonian toward the west. This submarine landslide complex may also have provided a lateral source for landslides through secondary remobilisation perpendicular to the regional palaeflow from the north. Bed pinch-out, thinning, ripple reflections and debrites provide further evidence for the presence of basinal topography during deposition.

The Coniacian-Maastrichtian mixed siliciclastic-calcareous deep marine system contains both siliciclastic and calcareous lobe elements, which represent different lobe sub-environments, requiring modification of terminology developed for siliciclastic lobes. Mixed systems are also shown to have unique facies, both in outcrop and a subsurface analogue from offshore The Gambia, reflecting differing depositional processes between the systems operating contemporaneously. Interaction between the two deep-marine environments characterising the mixed systems has also made stacking patterns difficult to decipher, with each system attenuating the other.

Acknowledgments

Table 1:Summary of observations and interpretations of lithofacies, used to create facies associations

Facies	Description	Interpretation
Conglomerates (A)	0.1 to 3 + m thick beds of poorly-sorted, disorganised conglomerates. Most commonly clast-supported consisting of sub-angular to sub-rounded boulder-, cobble- and pebble-sized clasts of limestone and sandstone. Matrix comprises a poorly-sorted mix of all finer size fractions. Cm- 10cms scale mud-clasts occur sporadically throughout the beds. Bed bases are often erosive, and can be amalgamated. This facies often grades into thick bedded sandstones (C).	The characteristics of this facies suggest deposition from debris flows having cohesive as well as frictional strength (Fisher, 1971; Nemeck & Steel, 1984). The grading of conglomerates into thick-bedded sandstones reflects the transition of hyperconcentrated submarine debris flows into highly-concentrated turbulent flows (Mulder & Alexander, 2001; Sohn et al. 2002), due to the entrainment of ambient water (Postma et al. 1988).
Poorly sorted clast rich deposit (B)	0.1 – 1+ m thick poorly sorted deformed, matrix-supported units. Matrix can range from mudstone to coarse sandstone, and is often poorly-sorted and sheared. Clasts include cm-m scale limestone and sandstone blocks, rafts of remobilised folded thin-bedded sandstones, sporadic pebbles and granules and frequent mud-clasts. These deposits are commonly non-graded, but can show weak normal-grading.	The poorly-sorted matrix and large clast sizes are suggestive of 'flow freezing' of a flow with yield strength (Inverson et al. 2010), indicating 'en masse' deposition from a laminar flow (Nardin et al. 1979; Inverson 1997; Sohn 2000). Remobilised thin-bedded sandstones and intra-basinal clasts indicate localised mass failure and reworking.
Thick-bedded sandstones (C)	0.5 – 1+ m brown siliciclastic fine-granular sandstones. Normally-graded or non-graded and typically lacking in primary depositional structures. Bases are often sharp and erosive. Parallel laminations are sometimes present at bed tops and mud-clasts can be observed throughout. Weak cross-lamination is infrequently observed.	The general massive structuration of these deposits suggests that they represent rapid aggradation beneath a highly concentrated but dominantly turbulent flow, and are thus interpreted as high density turbidites (Lowe, 1982; Mutti 1992; Kneller & Branney 1995).
Mixed siliciclastic and calcareous sandstones (D)	0.1-1m beds of medium-bedded calcareous sandstones with punctuated interbeds of cm-scale thin-bedded siliciclastic sandstone, either as continuous beds or lenses. The medium-bedded calcareous sandstones are massive and the siliciclastic beds are often erosively-based and show tractional structures (ripple and planar lamination). Siliciclastic beds can be amalgamated with each other or isolated between calcareous siltstones or sandstones.	Medium-bedded calcareous sandstones are interpreted to represent deposition from a slowly aggrading dilute turbidity current. Periodic, thin-bedded siliciclastic sandstones represent deposition from a relatively quickly aggrading dilute turbidity current, which interacted with a much slower aggrading calcareous turbidity current.
Medium-bedded calcareous sandstones (E)	0.1-1 m thick beige beds of calcareous siltstone -fine sandstone. Normally-graded or non-graded. Planar lamination may be present, but other tractional structures are rare. Beds can be amalgamated.	Based on their tractional structures and normal-grading, beds are interpreted as having being deposited from dilute turbidity currents. These beds are interpreted as medium-density turbidites, due to larger bed thickness and infrequent tractional structures, than thin-bedded calcareous sandstones (G). Thicker beds and fine grain size indicate a slowly aggrading dilute turbidity current.
Medium-bedded siliciclastic sandstones (F)	0.1 -0.5 m thick brown beds of very fine – granular grained, commonly normally-graded, sandstones. Inverse-grading is infrequently observed. Basal parts of the bed are often structureless containing infrequent cm-scale mud-clasts while tops are rich in tractional structures including parallel, ripple and hummock-like laminations. Bed bases are often erosive, and can be amalgamated.	Based on their tractional structures and normal-grading, beds of this lithofacies are interpreted as deposition from a dilute turbidity current. These beds are interpreted as medium-density turbidites due to their bed thickness and common lack of structures in the lower part of the bed (e.g. Soutter et al. 2019).

Thin-bedded calcareous sandstones (G)	0.01 – 0.1 m thick beige beds of calcareous siltstone-fine sandstones. Can be normally-graded, often into silty-mudstones, or not graded. Planar laminations are observed but other tractional structures are typically absent. Individual beds are often amalgamated.	Thin-beds, fine grain size and weak planar laminations represent deposition from a low-concentration turbidity current (Mutti 1992; Jobe et al. 2012; Talling et al. 2012), indicating these beds are low-density turbidites. Fine grain size, thicker beds compared to thin-bedded siliciclastic sandstone (H) and absence of ripple laminations suggest slowly aggrading, dilute remnants of a turbulent flow, (Remacha & Fernández 2003; Bell et al. 2018), which did not reach significant velocity to generate ripple laminations (Baas et al. 2016).
Thin-bedded siliciclastic sandstones (H)	0.005 – 0.1 m thick brown beds of siliciclastic very fine- granular sandstones. Commonly normally-graded, occasional inverse-graded. Tractional structures (planar, ripple, hummock-like and convolute laminations) and sporadic mud-clasts are observed. Bases can be flat or weakly erosive and sometimes contain granules. Bed tops are often flat. Where present, ripples can show opposing palaeoflow.	Thin-bedded, structured sandstones are interpreted to be deposited from low-concentration turbidity currents (Mutti 1992; Jobe et al. 2012; Talling et al. 2012) and are therefore interpreted as low-density turbidites. Ripples with opposing palaeoflow suggests topographic interference.
Bi or tri-partite beds (I)	0.05-0.5 m thick beds that contain multiple parts. Typically consisting of a lower fine-coarse sandstone (division 1) overlain by a poorly-sorted muddy siltstone – medium sandstone (division 2). Division 3 is sometimes present consisting of a siltstone-fine grained sandstone loaded into division 2. Divisions 1 and 3 sometimes contain planar laminations and sporadic cm scale mud-clasts. Division 2 is often highly deformed and rich in mud-clasts and very coarse sandstone to pebble-grade clasts.	Tractional structures in division 1 and 3 indicate formation under turbulent flows. Poor-sorting and mud content suggest division 2 was deposited under a transitional-laminar flow regime. These bi-tri partite beds are hybrid beds (Haughton et al. 2009), generated by flow transformation from turbulent to laminar. Such transformation occurs through flow deceleration (Barker et al. 2008; Patacci et al. 2014) and by an increase in concentration of fines during flow run-out (Kane et al. 2017).
Mudstone (J)	0.005 – 8 m thick pale grey or red mudstone – fine siltstone beds, which are friable and often inferred in areas of missing section. Planar laminations, discontinuous drapes and lenses of siltstone may be present. Commonly calcareous in composition. Red beds are common at the base of the Campanian.	Low energy conditions, representative of background sedimentation via suspension fallout. Laminations may be present below the scale visible in outcrop, representing deposition from a dilute turbidity current (Boulesteix et al. 2019). Pale colour indicates low total organic carbon (TOC). Red beds are similar to Cretaceous Oceanic Red Beds (CORBS) described across Europe (Wang et al. 2005; Hu et al. 2005; Wagreich & Krenmayr, 2005) and represent deposition below the carbonate compensation depth (CCD) in a deep oceanic basin.

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1138
1139 **FIGURES**
1140
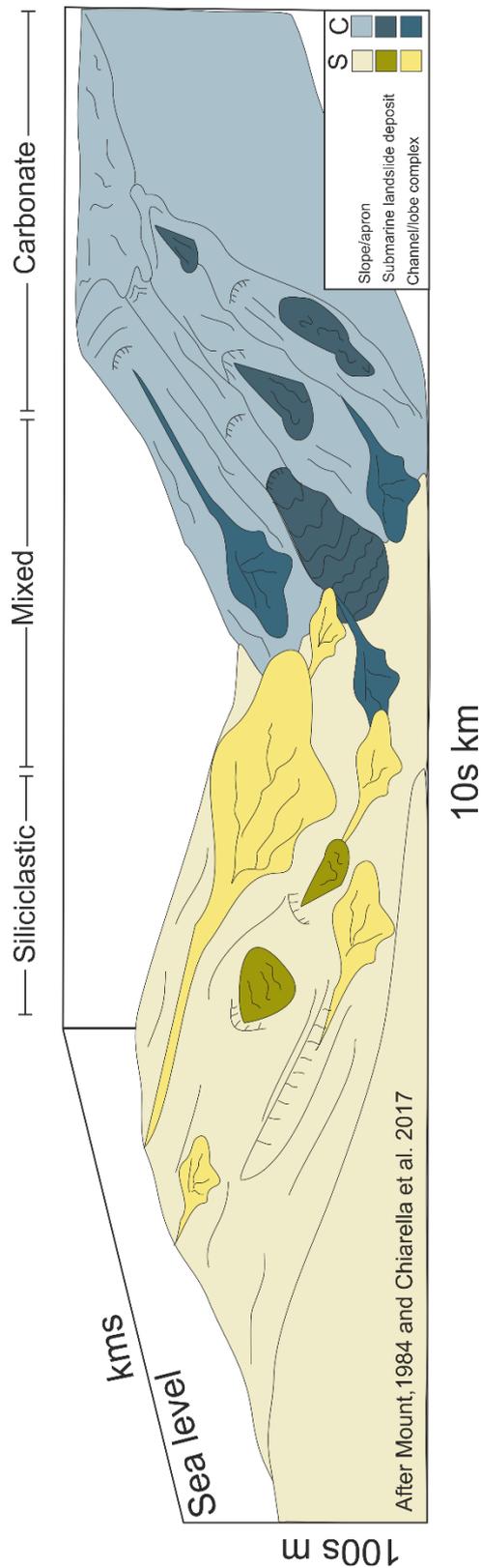
Figure 1: Simplified conceptual model showing how siliciclastic and carbonate systems may interact at a basin-scale in a deep-marine mixed siliciclastic-carbonate system. Carbonate material is shed from a shallower carbonate-producing platform that periodically received siliciclastic material; this is then redeposited in the deep-marine by of gravity flows (After Mount 1984, Chiarella et al. 2017).

A

48°E

Russia

Lithology

N	Plio-Pleistocene Miocene
P	Paleogene
K₂	Upper Cretaceous
K₁	Lower Cretaceous
J₃	Upper Jurassic
J₂	Middle Jurassic
	Olistromes

Structure (map)
Major reverse fault
Major normal fault

Structure (cross-sections)
Faults
Faults (inferred)
Bedding
Axial planes

N — 10 km

Qusar
Quba
Tahircal
Sudur
Sahdag
Qizilqaya
Xinaliq
Hapit
Cek
Buduq
Sohub
Ruk
Qarabulaq
Babadag
Yerfi
Kelvudag
Qonaqkend
Tangaalti

Study area
Tufan
Figure 3
41° N

P, K₁, K₂, J₂, J₃, N, B, C

B

NE Tahircal Zone — Tufan Zone SW

Cek Village
Hapit Village
Study area

4 km, 3 km, 2 km, 1 km, 0 km
0 km, 5 km, 10 km, 15 km

J₃, K₂, K₁, J₂

C

NE Tahircal Zone — Sudur Zone — Tufan Zone SW

Sahdar - Besbarmag N. (top) - Sahdag - Xizi Z. (bottom)

Tahircal Village
Sudur Village
Sahdag Mt. (4243 m)
Bazarduzu Mt. (4466 m)
~Study area

5 km, 4 km, 3 km, 2 km, 1 km, 0 km
0 km, 5 km, 10 km, 15 km, 20 km, 25 km, 30 km, 35 km

N, J₂, J₃, K₁, K₂, P

Figure 2: Structural and stratigraphic framework of the Eastern Greater Caucasus (EGC) of
Azerbaijan. A) Simplified geological map, black box locates study area, lines B and C are located.
Schematic location of Figure 3 is shown. B) Cross section across the Cek-Hapit Valley, black box
locates study area. C) Cross section from Tahircal-Sahdag Mountain – Bazarduzu Mountain,
black box locates equivalent stratigraphy to our studied section (Modified from Bochud, 2011).
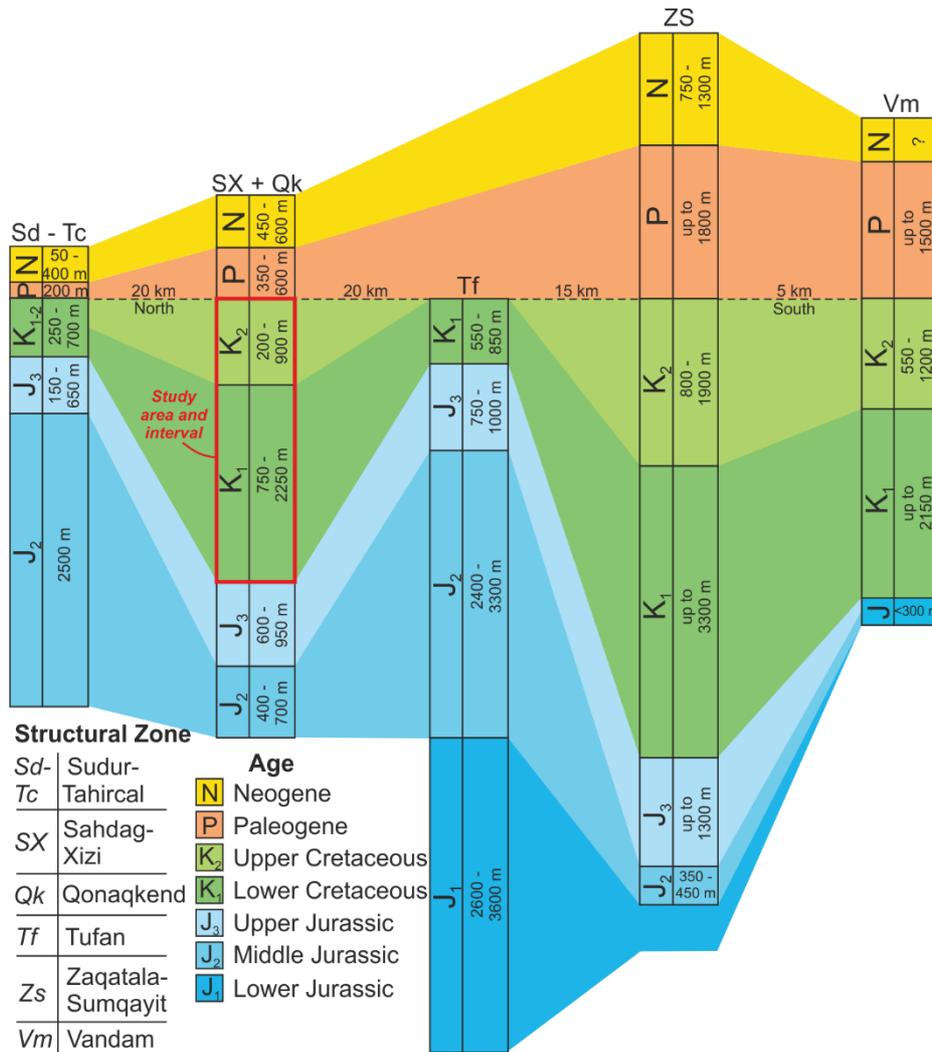
Figure 3: Stratigraphic section trending roughly north-south across the five main structural zones
(from Bochud 2011) of the EGC. Flattened along the top of the Cretaceous, located on Figure 2.

Cretaceous Structure of the Qonaqkend Zone

Legend: LC (N = 17), UC (Cek) (N=58), UC (Qonaqkend) (N=15)
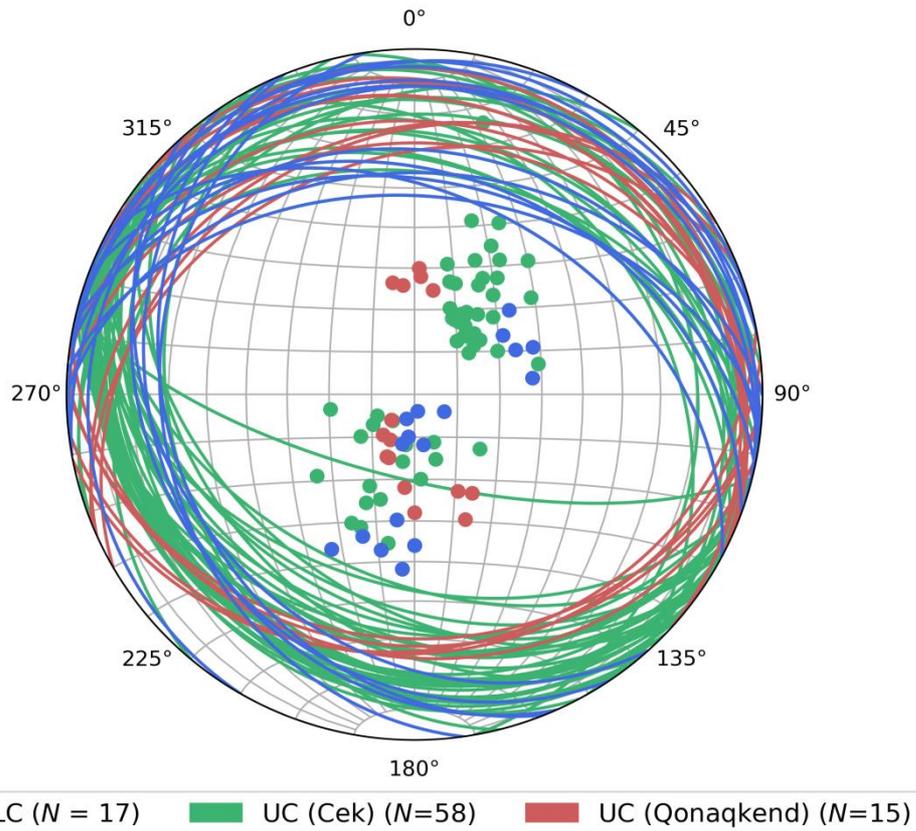
1162
1163 Figure 4: Equal area stereographic projection showing bedding readings for Cretaceous
1164 stratigraphy across Qonaqkend Zone. Bedding planes shown as lines and poles to bedding shown
1165 as dots. Coloured by stratigraphy and location; LC- Lower Cretaceous, UC- Upper Cretaceous.
1166 Structural data reveals a shallow-moderate structural dip to the north and south, in agreement with
1167 the east-west trending structural zones of the EGC.

Cenomanian-Turonian Paleocurrents (Qonaqkend) N = 18

Conacian-Santonian Paleocurrents (Qonaqkend) N = 5

Campanian-Maastrichtian Paleocurrents (Cek) N = 33
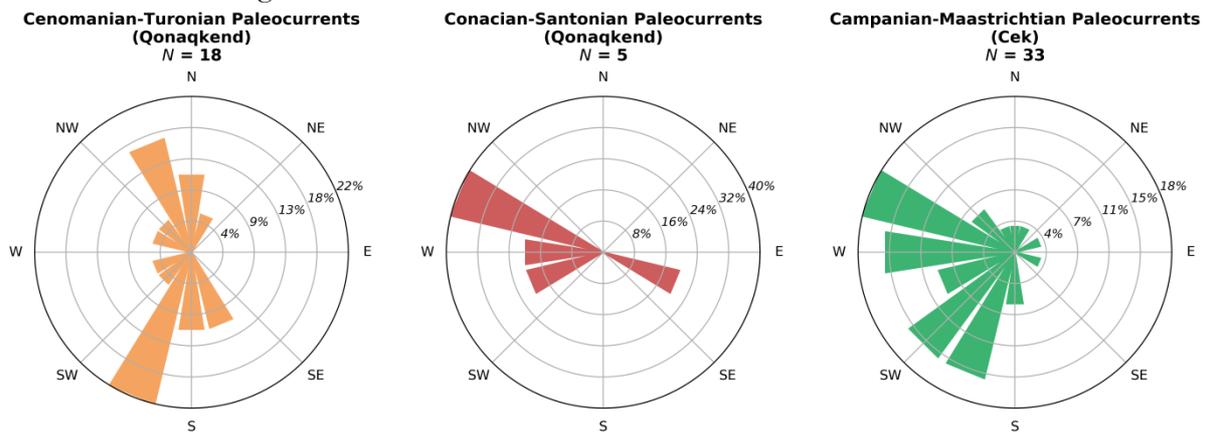
1168
1169 Figure 5: Rose diagrams from palaeocurrent indicators (ripples, sole marks, cross-stratification)
1170 from the Cretaceous stratigraphy of the Qonaqkend Zone. Readings have been corrected for
1171 tectonic tilt and are subdivided by stratigraphy and location (see Figure 2).

1172
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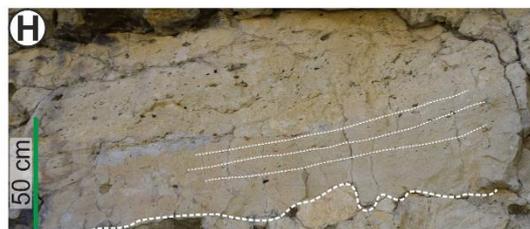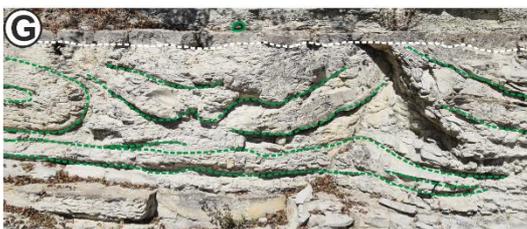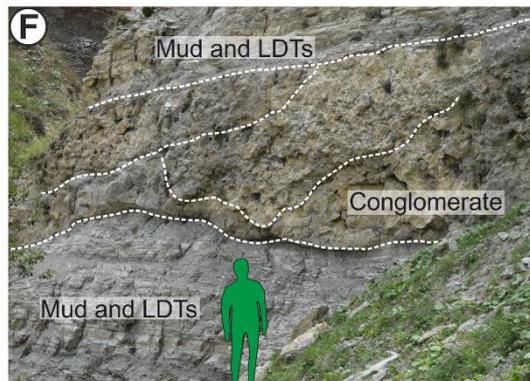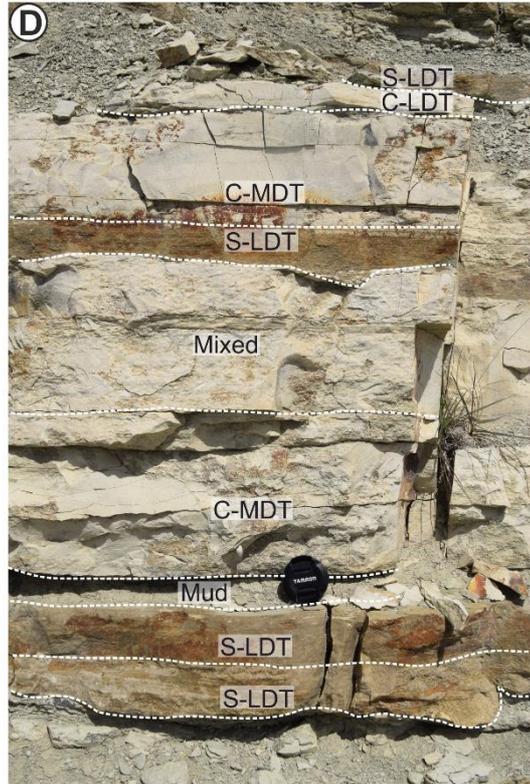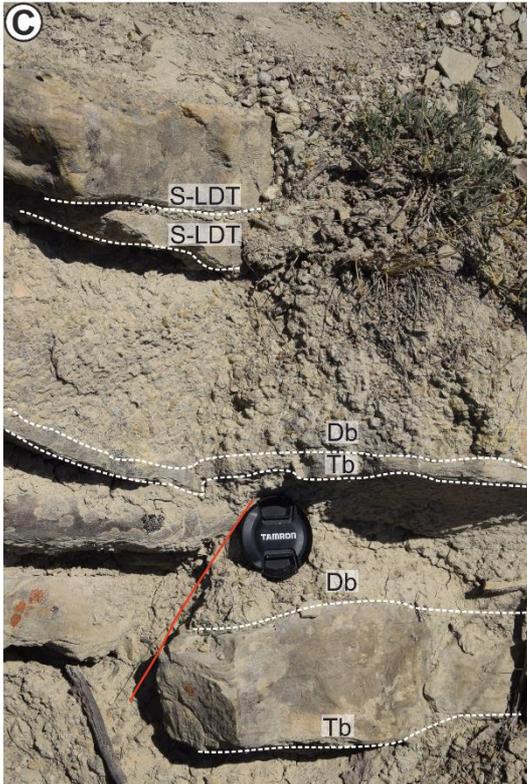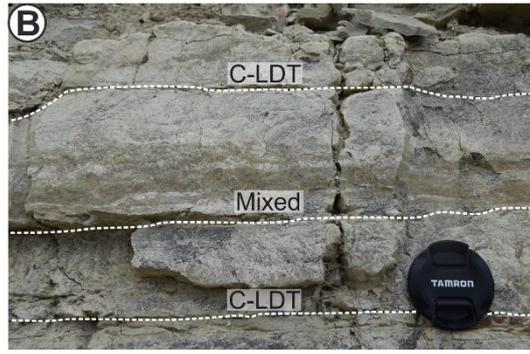
1175   Figure 6: Facies photographs. Facies described in detail in Table 1. Scale is either lens cap (52
1176   mm), person (1.74 m) or indicated. LDT; low density turbidite, MDT; medium density turbidite,
1177   Db; debrite (poorly sorted clast rich deposit); Tb; Turbidite, S; Siliciclastic, C; Calcareous. A)
1178   Calcareous mudstone B) Calcareous low density turbidite and mixed beds (of siliciclastic and
1179   calcareous low density turbidites). C) Two bi-partite beds consisting of a lower turbidite and an
1180   upper debrite, in this case both siliciclastic, overlain by two siliciclastic low density turbidites. D)
1181   Evidence for facies scale mixing (sensu Chiarella et al. 2017); calcareous turbidites were
1182   recognised in the field by their pale cream colour, while siliciclastic turbidites were brown-orange
1183   in colour and contained visual quartz granules. Calcareous turbidite probably accumulated slowly
1184   based on their grain size, and were punctuated by siliciclastic gravity flows, forming mixed beds.
1185   E) Siliciclastic low and medium density turbidites with cm-scale mud clasts weathered out. F)
1186   Mudstone and low density turbidites (both calcareous and siliciclastic) punctuated by metre-scale
1187   amalgamated conglomerates. G) Chaotic, clast rich-deposit with deformed, non-extensive
1188   bedding. Camera lens cap circled in green. H) Erosionally-based, crudely cross-laminated
1189   siliciclastic high density turbidite rich in mud clasts.

1190

Figure 7: Type examples of the seven recognised facies associations, divided into siliciclastic and mixed (siliciclastic and calcareous) associations, by orange and blue boxes respectively. Scale either lens cap (52 mm), person (1.74 m) or indicated. 10 m type log is taken from representative logged section of each facies association. Cretaceous Oceanic Red Beds; CORBS.

Coniacian-Maastrichtian Bed Statistics
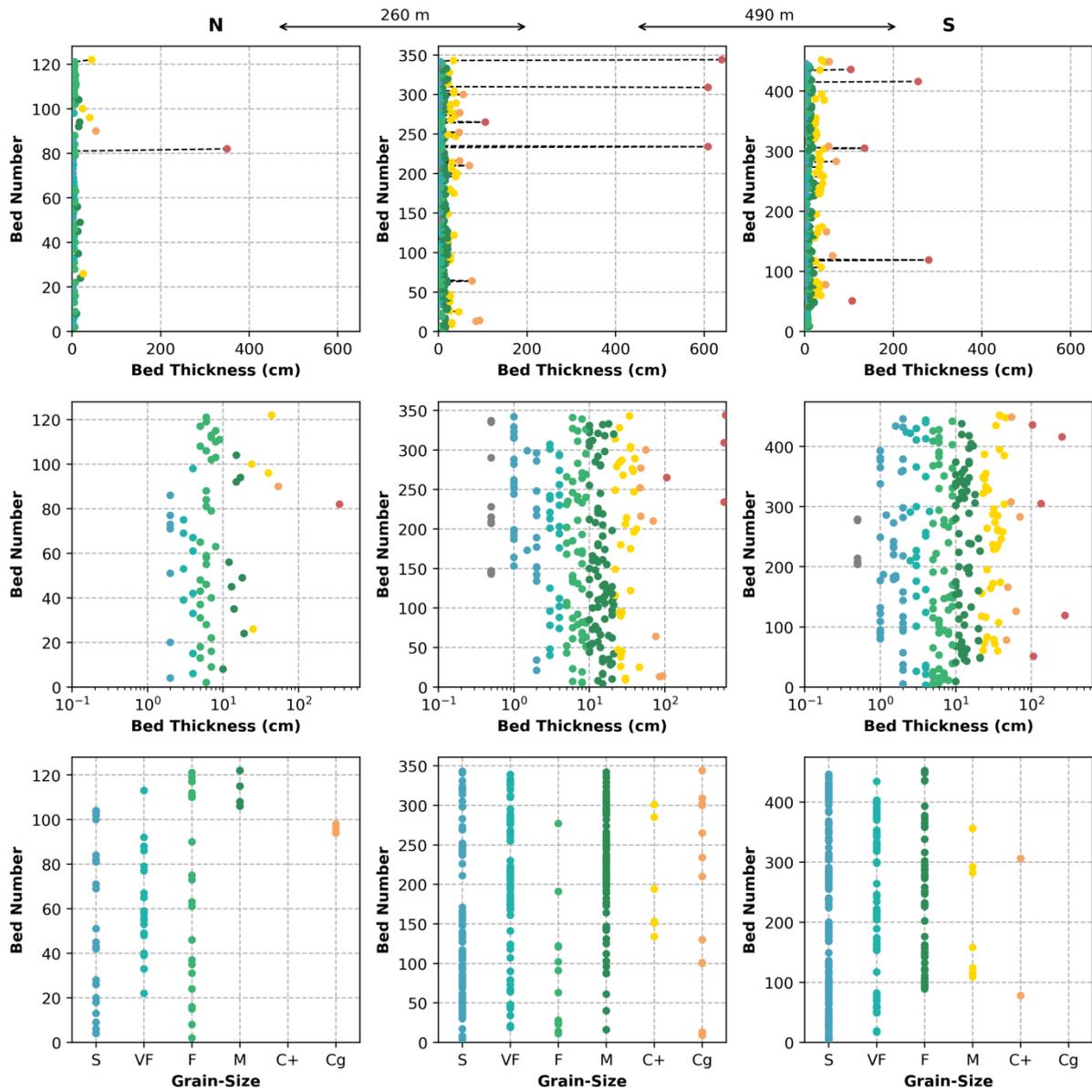
Figure 8: Quantative facies analysis for Coniacian-Maastrichtian stratigraphy. The three columns represent three different logged sections from north to south that are representative of northern margin, axis and southern margin of the basin respectively. Charts compare bed number (with 1 being at the base of the log and 200 at the top) to bed thickness (linearly in the top column and logarithmically in the middle column) and logged grain size (in the basal column). Where grain size varies within the bed average grain size is used. In the top column thick beds are highlighted with a dashed line. Colours are for visual separation of data, and meaning changes per column, blue are thin/fine beds/grain-size, oranges are thick/coarse beds/grain-sizes. Scales for bed number vary across the rows.
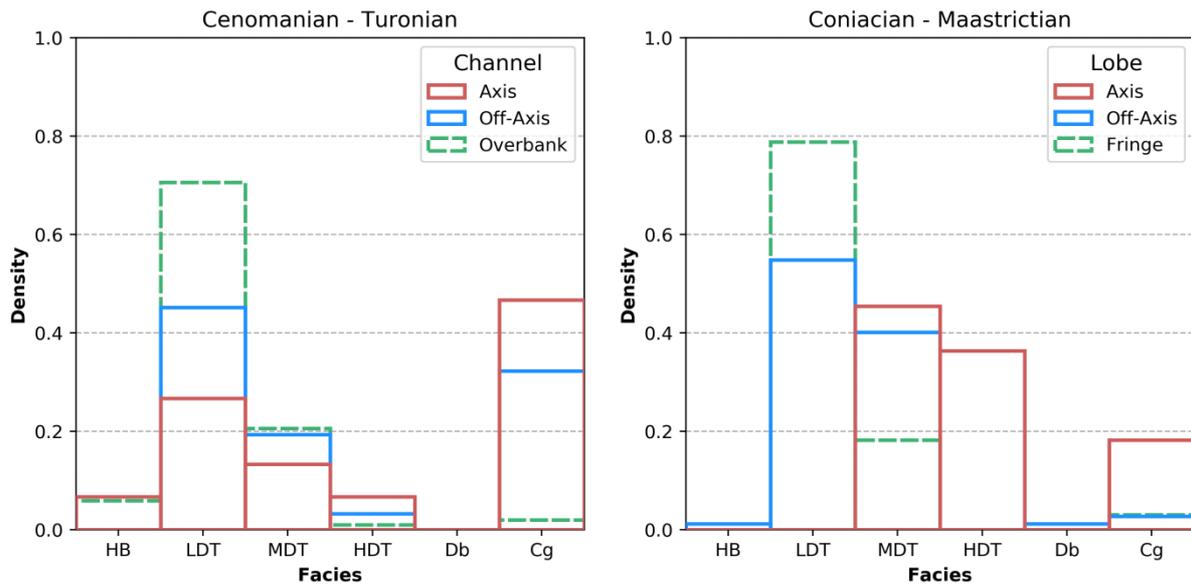
Figure 9: Quantative facies analysis for Upper Cretaceous stratigraphy, divided into Cenomanian-Turonian channelised siliciclastic deposition and Coniacian – Maastrictian mixed lobe deposition. Facies refer to HB: hybrid bed (bi and tri-partite beds); LDT: low density turbidite; MDT: medium density turbidite; HDT: high density turbidite; Db: debrite (poorly sorted clast rich deposit) and Cg: conglomerate. Coloured by type log for different sub-environment. Density refers to the percentage of each facies in each log.
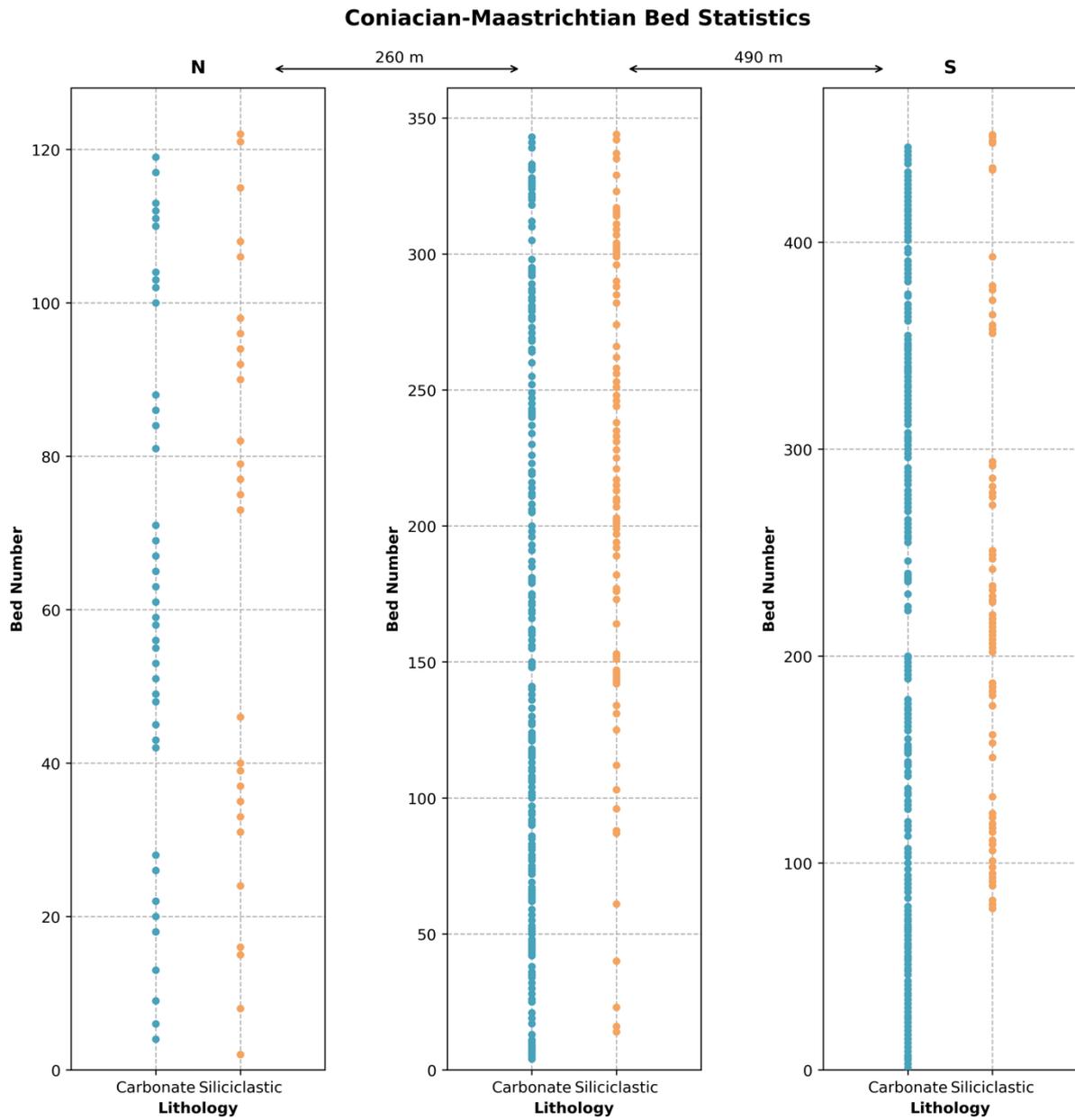
Figure 10: Quantative facies analysis of Coniacian-Maastrichtian mixed stratigraphy comparing bed composition (carbonate or siliciclastic) against bed number (height up log). Using the same logged sections as Figure 8, and thus a different number of beds per log resulting in variable bed number scale.
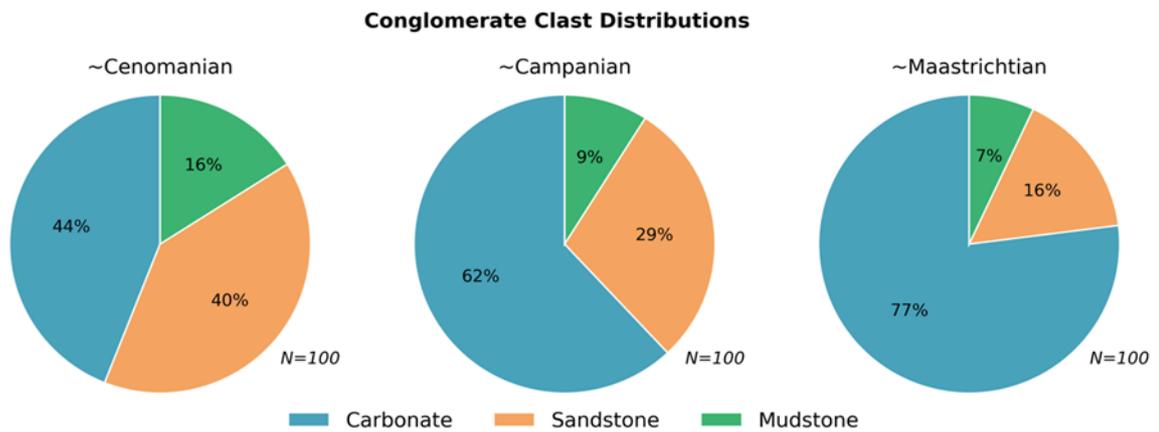
1218   Figure 11: Pie charts showing composition of clasts within conglomerates per stratigraphic age, taken
1219   from 100 clasts from representative conglomeratic beds of over 1 metre thick. Percentage equates to
1220   absolute number of clasts, as 100 are sampled. Carbonate clast content increases through time, discussed
1221   in text.

1222

Figure panels:

B — 5 cm

C — Conglomerate onlap. LDT running up topography.

E — Slumping to the south. 5 m

A — Slope angle reconstruction. 6°. 32°−26° = 6°. Campanian 26°. Santonian 32°. Jurassic.

D — Isolated channelised conglomerate. Downward pinch out. Upward pinch out.

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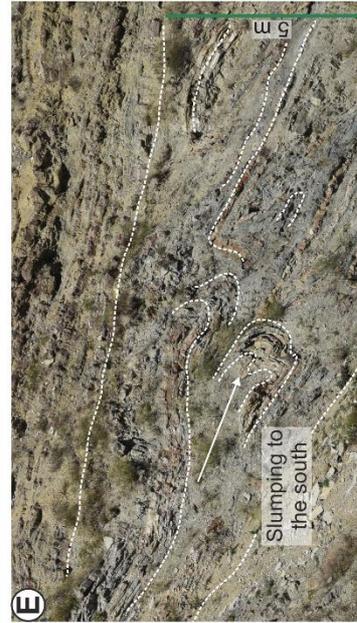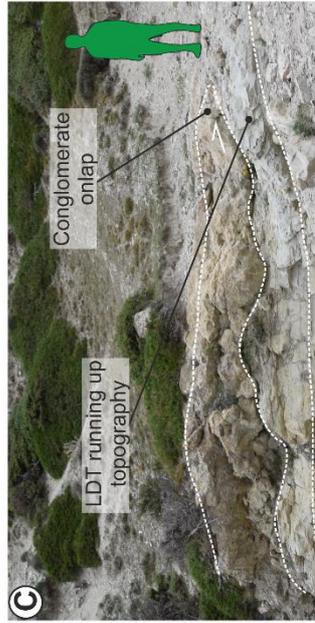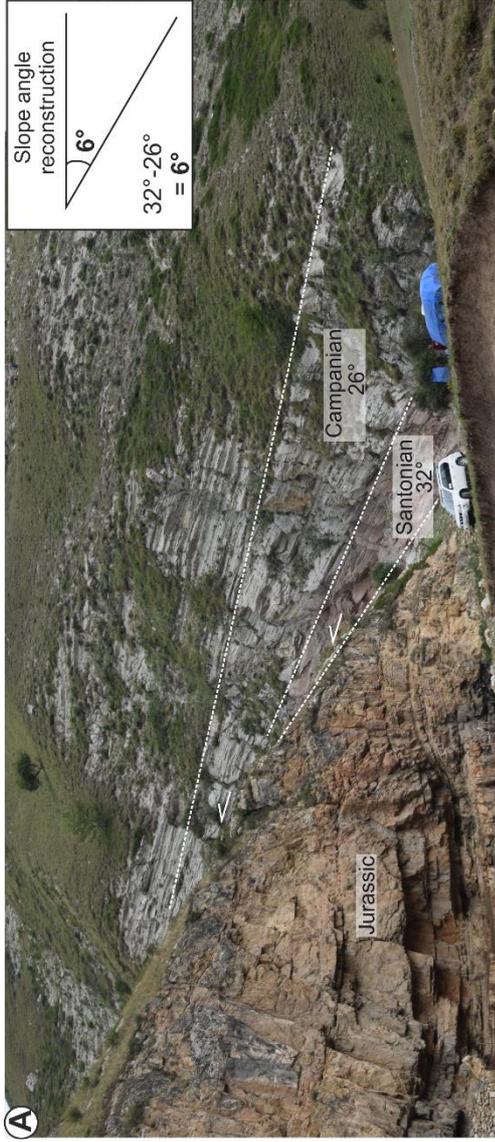
1224   Figure 12: Evidence for palaeotopography. Scale indicated by person (1.74m), car (1.9 m) or
1225   indicated. A) Cretaceous stratigraphy thinning and onlapping Jurassic limestone, slope angle
1226   reconstructed. B) Evidence for opposing ripple directions suggesting flow reflection. C)
1227   Thickness and pinch-out variability of different deposits on a metre-scale laterally. D) Cliff
1228   section containing three conglomerate bodies which vary in architecture and termination style as
1229   indicated. E) Submarine landslide deposit showing evidence for slumping towards the south.

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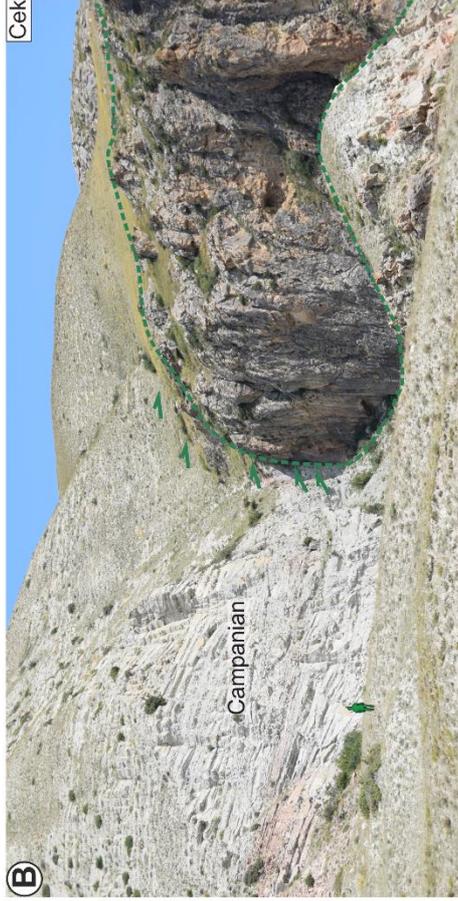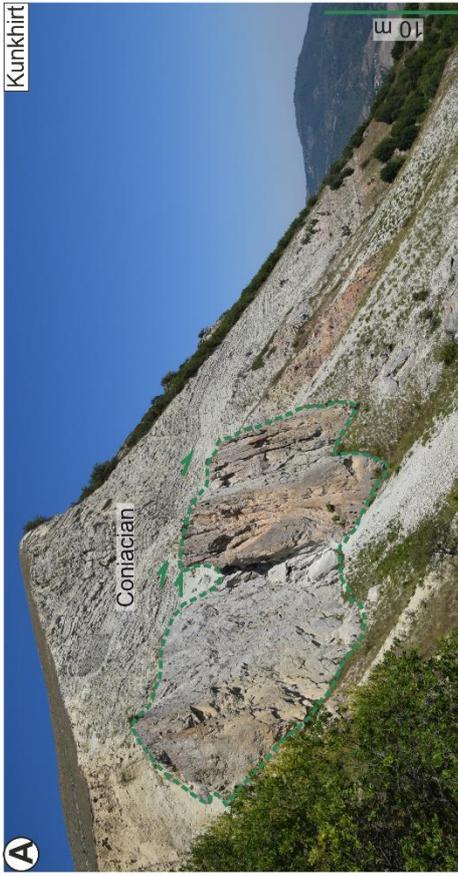
Figure 13: Evidence for collapse of the Jurassic carbonate platform. Scale is shown by person (1.74 m), car (1.9 m) or indicated. Onlap indicated by green arrows, extent of clast or bedding surface shown in green dashed line. A) Metre-scale limestone clasts within Coniacian mixed stratigraphy at Kunkhirt. B) Campanian calcareous low density turbidites abutting against a decametre-scale Jurassic limestone block at Cek. C) Stacked, inversely-graded submarine landslide deposits primarily composed of reworked Jurassic limestone blocks at Rük, upwards widening triangles indicate coarsening upwards (inverse-grading). D) In situ break-up of the Jurassic platform at Cek. Clasts are fractured and separated but bedding planes (indicated in white) are still visible showing minimal displacement.
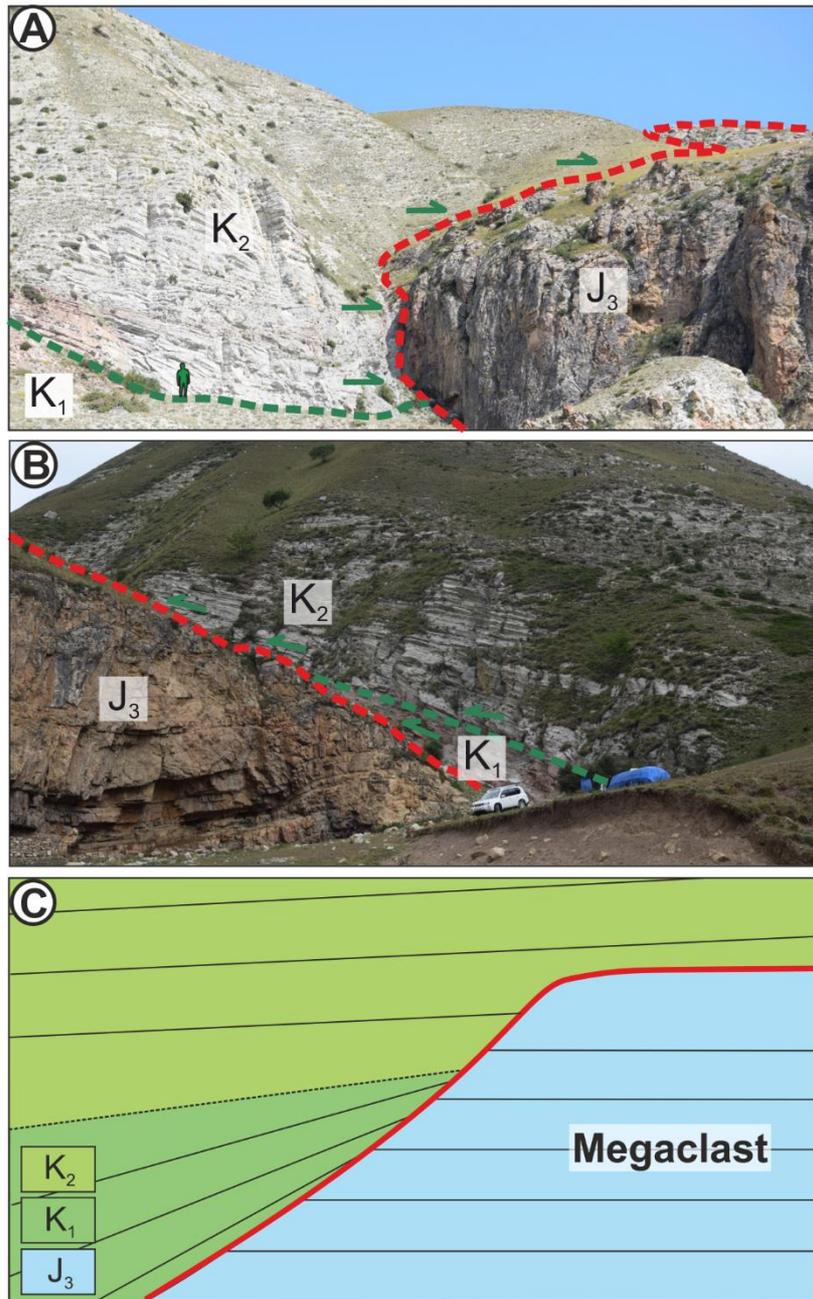
Figure 14: Evidence and model for the generation of topography by an allochtonous block throughout the Cretaceous. A&B) field examples of Jurassic stratigraphy forming topography throughout the Cretaceous influencing sediment routing. C) Schematic model showing formation of topography.
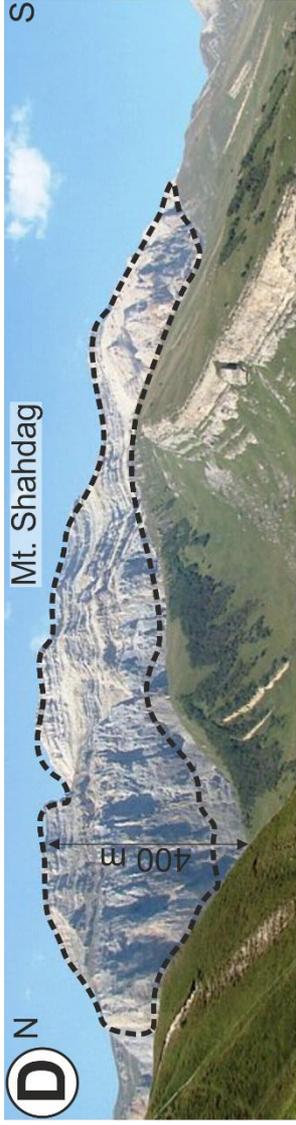
1244   Figure 15: Evidence for allochtonous block model as the most likely for the generation of
1245   Cretaceous topography. Scale is shown by person (1.74 m) or indicated. Blocks are drawn
1246   around with black dashed line. A-C are examples of metre-decametre scale clasts around Cek. D)
1247   Shahdag Mountain, the tallest mountain in Azerbaijan, is interpreted as a kilometre-scale wide
1248   olistostrome by Bochud, 2011. E) Block with both margins exposed and onlapped by mixed
1249   stratigraphy, black box locates C.

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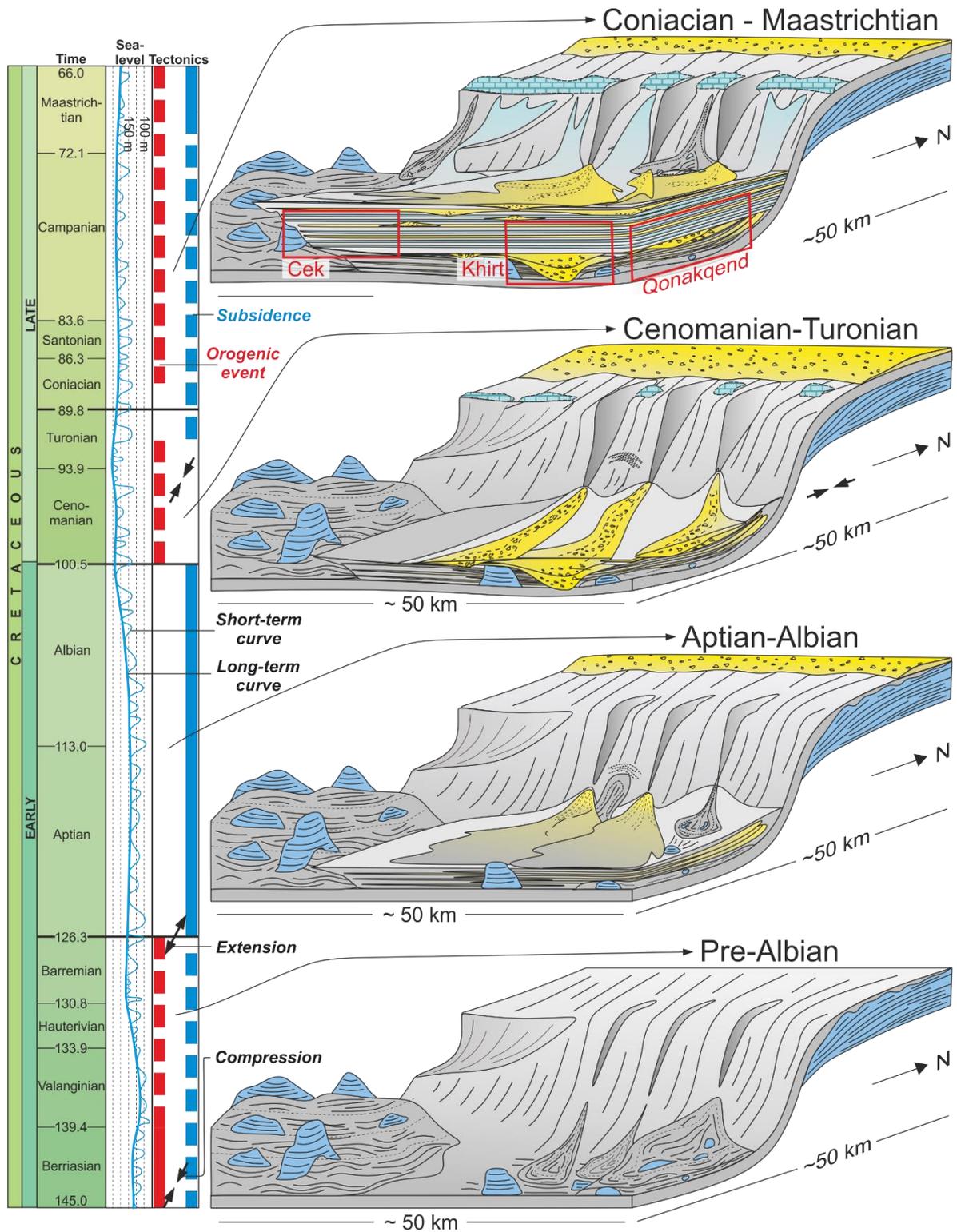
Figure 16: Evolutionary model for the Cretaceous of the study area. Studied stratigraphic sections highlighted. Topography, thought to be formed by a mega-clast, is present throughout the Cretaceous and influences deposition, discussed in text. Extract from the geological time column, sea level fluctuations and local tectonic events highlighted on the left. The Pre-Albian was dominated by limestone blocks on a muddy slope. Thin-bedded siliciclastics of a distal lobe were deposited during the Aptian-Albian. Siliciclastic channels are prominent throughout the Cenomanian-Turonian. In the Coniacian-Maastrichtian mixed calcareous and siliciclastic lobes, of different sub-environments interact, and are likely sourced from the same northern margin.

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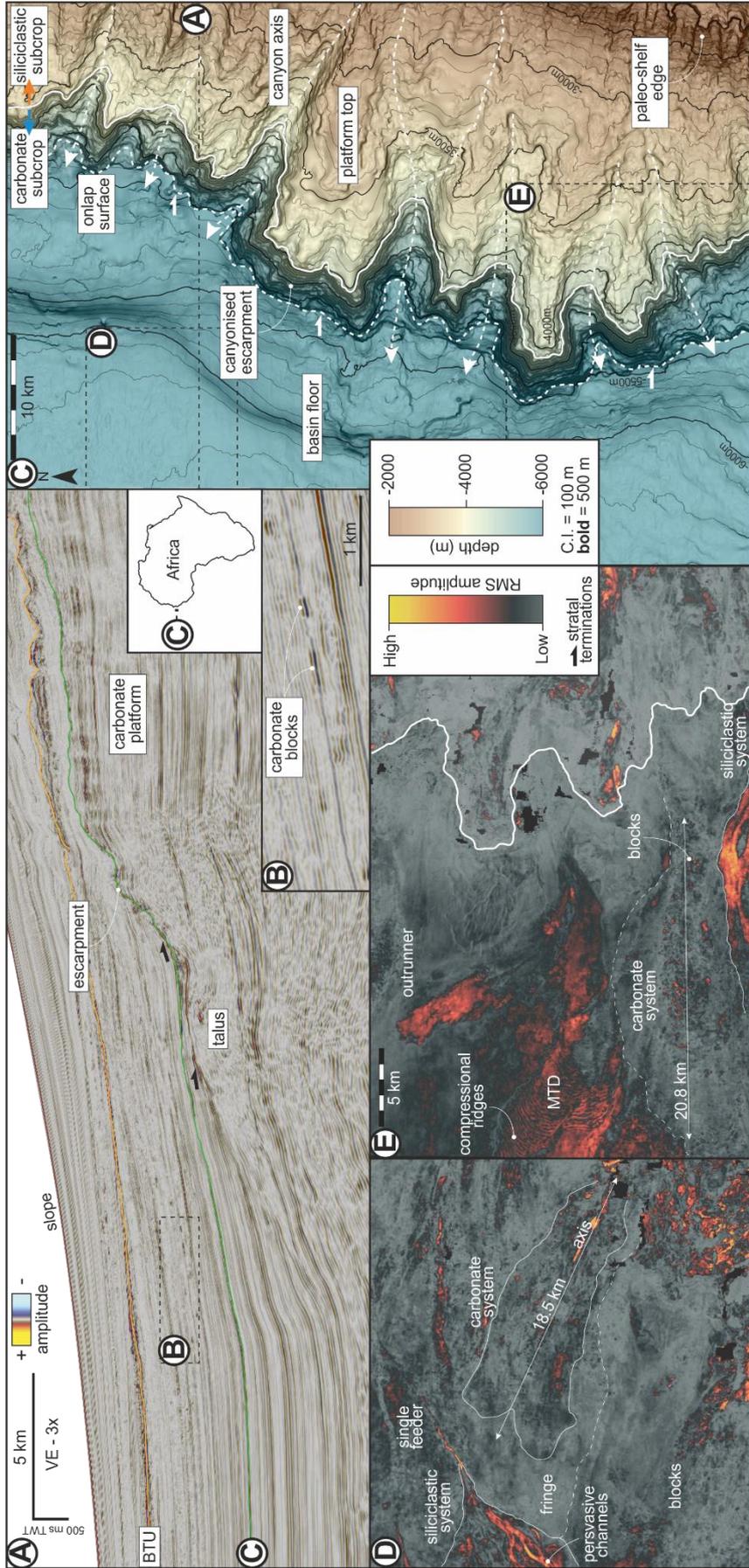
1262 Figure 17: Seismic geomorphology of a mixed-system offshore The Gambia, NW Africa. A)
1263 East-West two-way time seismic section with key stratigraphic surfaces, location shown in C. B)
1264 Enlarged two-way time seismic section of carbonate blocks observed in the deep-marine mixed-
1265 system. C) Depth structure map (contoured) with dip magnitude below of the diachronous
1266 regional composite unconformity surface displayed on seismic section A, showing the heavily
1267 canyonised carbonate escarpment margin, canyon flow pathways (dashed white lines),
1268 lithological contact between siliciclastic and carbonate subcrop (solid white line) and onlap
1269 surface (dotted white line). D & E) RMS amplitude maps extracted from a +/- 12 ms window
1270 around two Late Cretaceous horizons in the basin. Location of maps shown on C.

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Carbonate/Siliciclastic Thickness & Grain-Size Trends

1277

51
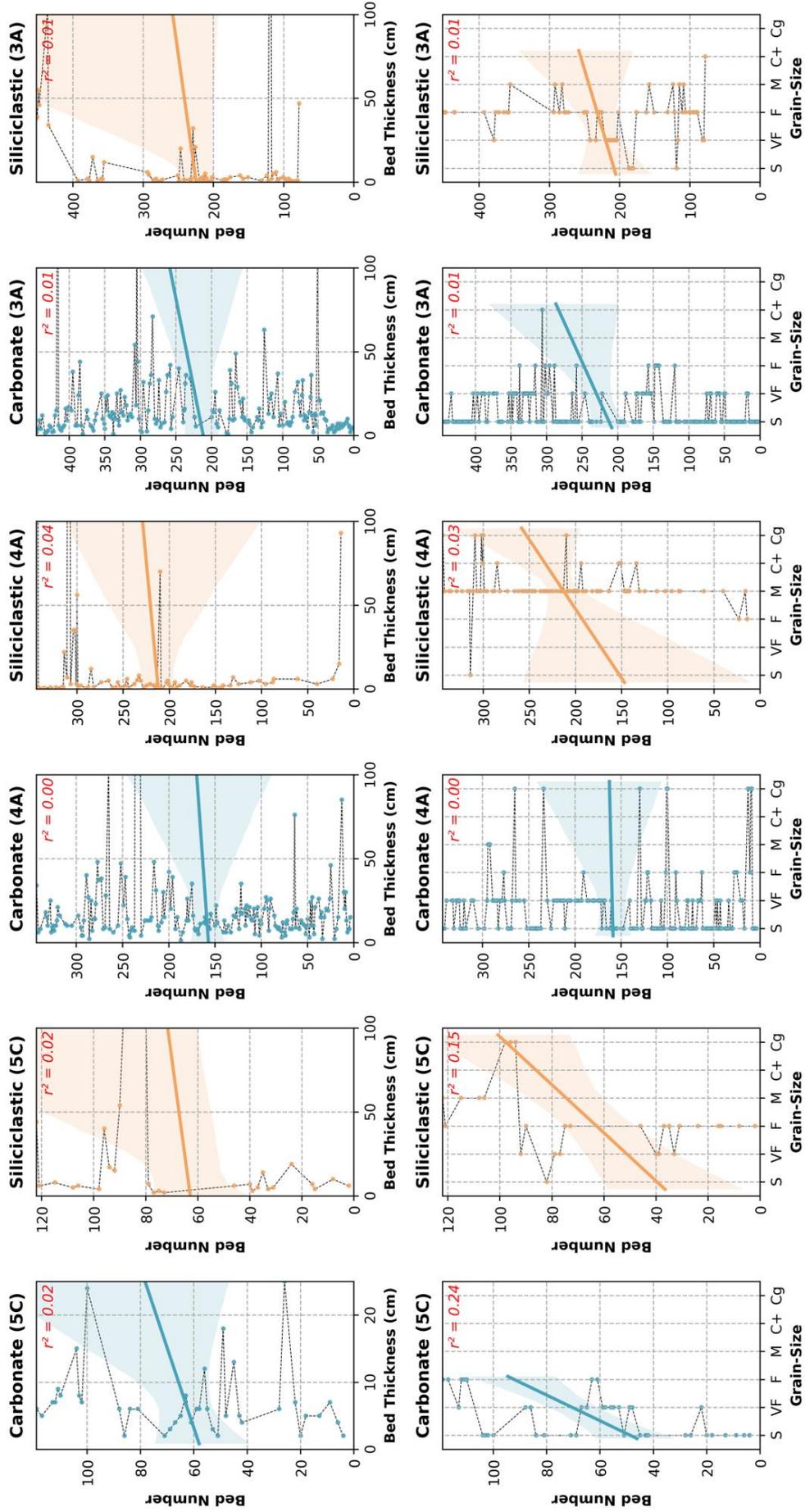
1278   Figure 18: Quantitative facies analysis for Coniacian-Maastrichtian stratigraphy. Logs are the
1279   same as Figures 8 and 10. 5C, 4A and 3A represent northern margin, axis and southern margin
1280   respectively. Each log is divided into carbonate and siliciclastic beds, and bed number is plotted
1281   against bed thickness (upper row) and grain size (bottom row). Lines indicate linear regression, r^2
1282   value is labelled in red, shading either side of the line indicates a 95% confidence interval for the
1283   linear regression (trend line). Low-no correlation is observed suggesting no stacking patterns or
1284   modulation of any stacking patterns by interaction of both systems, discussed in text.
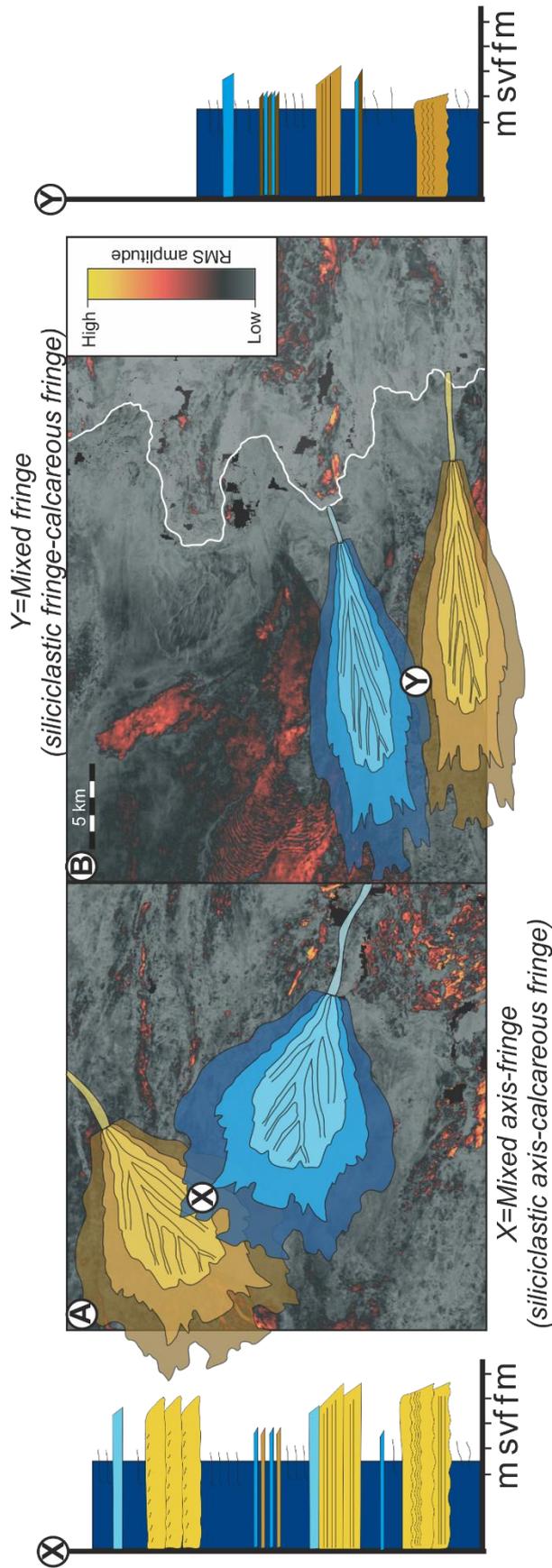
Figure 19: Schematic showing potential interactions of calcareous and siliciclastic lobes in mixed
systems. A and B are RMS maps from Figure 17, which have been overlain by lobe complex

1289 geometries, as an interpretation based on seismic facies analysis and understanding of regional
1290 source area (see Casson et al. 2020). X and Y represent log/core through locations where the
1291 lobe complexes interact in A and B respectively. X crosses the lobe fringe of the calcareous
1292 system and the lobe axis of the siliciclastic system and Y crosses the lobe fringe of both systems
1293 resulting in a thinner and finer grained succession when compared to X. This variability
1294 highlights difficulties arising from exporting sub-environment terminology developed in
1295 siliciclastic systems (e.g. Prélat et al. 2009) into mixed systems
1296