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Mid-to Late Holocene landscape changes in the Rioni Delta area (Kolkheti lowlands, W Georgia)

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1 **Mid- to Late Holocene landscape changes in the Rioni Delta area (Kolkheti lowlands, W**
2 **Georgia)**

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11

12 **Abstract**

13 The Kolkheti lowlands (Colchis, Colchian plain) form the central part of the extensive coastal lowlands
14 along the Black Sea coast of Georgia. Situated between the Greater and the Lesser Caucasus,
15 favourable climatic conditions resulted in a constant human occupation of the region during the
16 Holocene. However, due to continued deltaic sedimentation and alluviation of the river Rioni, the
17 configuration and the environmental conditions of the coast and its hinterland have changed
18 considerably; this was related to sea-level fluctuations of the Black Sea and variation of the sediment
19 supply. This study presents new data on the Holocene coastal evolution of Western Georgia. Based on
20 the geochemical and sedimentological analysis of sediment cores and trenches from the northern
21 part of the Kolkheti lowlands, between the Black Sea and the rivers Rioni and Khobistsqali, and a
22 robust chronology (^{14}C and IRSL dating), our goals are (i) to document the chronostratigraphy along
23 two coring transects; (ii) to decipher geographical and environmental changes along Georgia's Black
24 Sea coast; and (iii) to trace the sea-level evolution of the study area. Based on the succession of eight
25 facies, representing different depositional environments, our results suggest that significant
26 environmental changes took place throughout the last eight millennia. At least since 5000 cal BC, the
27 sedimentary record indicates the widespread existence of shallow lagoons. Floodplain-related fine-
28 grained alluvium accumulated on top of the lagoonal stratum. The progradation of the delta plain
29 between 3500 and 1500 cal BC was accompanied by the evolution of extensive swamps with peat
30 formation. The data indicate a gradual and moderate sea-level rise since ~6000 BC. Ultimately, this
31 and follow-up studies may provide a valuable background for the understanding of the
32 palaeogeographical context of ancient settlements in the area.

33

34

35 **1 Introduction**

36 With an area of ca. 412,000 km² and a maximum depth of 2,212 m the Black Sea is the largest anoxic
37 water basin in the world (Degens and Ross, 1972). It is connected to the Mediterranean via narrow
38 and shallow straits (Bosporus, Dardanelles). Over the last millennia, the Black Sea has been subject to
39 substantial environmental changes. In effect, from a giant freshwater lake during the Last Glacial
40 Maximum (Panin and Popescu, 2007) it evolved to its present state of salinity and sea level. While
41 some authors have evoked for remarkable oscillations of the postglacial sea-level rise (e.g. Ryan et
42 al., 1997, 2003; Chepalyga, 2007; Yanko-Hombach et al., 2007; Balabanov, 2007), a number of follow-
43 up studies reasonably challenged this interpretation by presenting new data on the Holocene sea-
44 level evolution (cf. Brückner et al., 2010; Fouache et al., 2012; Kelterbaum et al., 2012). Recent
45 studies have particularly focused on the northern and western parts of the Pontic region (e.g. Panin,

46 2007; Carozza et al., 2011; Cordova et al., 2011). In contrast, the eastern Pontic region has been
47 neglected so far, although extensive lowlands exist, for instance along the Black Sea coast of Georgia.
48 Here, protected between the Greater and the Lesser Caucasus, human occupation is evidenced in the
49 so-called Colchian plain (Colchis) at least since the Late Chalcolithic and Early Bronze Age
50 (Lordkipanidze, 1991; Papuashvili, 2002). In addition, as part of ancient Colchis, the famous Greek
51 colony of Phasis is assumed to be located in the area of the present Rioni delta (Gamkrelidze, 1992,
52 2012; Lordkipanidze, 2000; Korenjak, 2003), although remnants of the city have hitherto not been
53 discovered, and its exact position unknown (Lordkipanidze, 1991; Gamkrelidze, 2012). The lack of
54 studies focussing on the coastal evolution, sea-level and environmental changes, particularly against
55 the background of the continuous occupational history in Western Georgia (Gamkrelidze, 2012),
56 demands an intensification of research activities in this area.

57 This paper contributes to fill knowledge gaps by adding new information on Holocene environmental
58 changes in the Kolkheti lowlands, which constitute the central part of the Colchian plain. Based on
59 eleven sediment cores, two sediment trenches, detailed sedimentological and geochemical analyses,
60 as well as radiocarbon and infrared stimulated luminescence (IRSL) dating, our study presents the
61 first systematic geoscientific investigation of the said coastal lowlands. More specifically, we aim at
62 (i) documenting the chronostratigraphy between the Rioni and Khobistsqali rivers along two coring
63 transects; (ii) deciphering palaeogeographical and palaeoenvironmental changes along the Georgian
64 Black Sea coast and its hinterland; and (iii) reconstructing the sea-level evolution for the study area,
65 and comparing it to other regional studies. Finally, this study represents a valuable background for
66 the interpretation of (future) local to regional archaeological studies, e.g. on the identification of
67 appropriate locations for the 'lost city' of Phasis.

68

69 **2 Regional setting**

70 *2.1 Geological and geomorphological framework*

71 The triangle-shaped Colchian plain (Fig. 1) is framed by the Black Sea in the west, the Greater
72 Caucasus in the north and the Lesser Caucasus in the south. Both mountain ranges formed during the
73 Alpine orogeny as a consequence of continent-continent-collision starting in the middle Pliocene (e.g.
74 Adamia et al., 2011). They consist of Mesozoic to Neogene igneous and sedimentary rocks and,
75 especially in the Lesser Caucasus, of Miocene to Quaternary volcanic rocks. In the east, where both
76 mountain ranges meet, the Surami range (or Likhi range) separates the Colchian plain from the Kura
77 basin, which drains into the Caspian Sea. The geologic basement of the Colchian plain mainly consists
78 of Cretaceous and Palaeogene sediments and volcanoclastics (Bazhenov and Burtman, 2002). These
79 strata are overlain by Quaternary strata, which were deposited by fluvial and marine processes
80 (Adamia et al., 2011). The Colchian plain is influenced by tectonic subsidence. Gamkrelidze (1998)
81 estimates subsidence rates of 2-4 mm/a for the central Colchian plain and 5-6 mm/a for the area
82 around Poti near the outlet of the Rioni.

83 While foothills of both mountain ranges reach the Black Sea coast south of Kobuleti and north of
84 Ochamchire, respectively, the floodplain of the Rioni dominates the Kolkheti lowlands. With a
85 catchment area of 13,400 km^2 it is the largest river of Western Georgia. Most of its tributaries
86 originate in the Greater Caucasus and dissect extensive foothills before entering the coastal
87 lowlands. Here, expanded wetlands with peat bogs, ponds, open reed areas and low forests
88 dominate; agricultural activities are impeded by the high water-table position. However, drainage

89   measures and ridge and furrow systems enable agriculture between and along the two rivers
90   (Nikolaishvili et al., 2015).

91   At the coast, the Rioni has formed a wave-dominated delta (Fig. 1), where longshore drift rapidly
92   redistributes the mostly sandy sediments along the graded shoreline. The flat and wide beaches are
93   backed by beach-foredune ridges, which are occasionally covered by pine forests. The main estuary
94   of the Rioni, formerly debouching into the sea directly in Poti, was relocated to its present position
95   during the Soviet era.

96   *2.2 Palaeoenvironmental changes and sea-level fluctuations*

97   Even though a number of studies on the Holocene landscape evolution exist from the second half of
98   the 20th century (e.g. Janelidze, 1980), most data from the Soviet era is difficult to access. During the
99   last 25 years only a few publications provide valuable information on Holocene environmental
100 development, with main focus on the vegetation history of Georgia's Black Sea coast since the mid-
101 Holocene (e.g. Connor et al., 2007; de Klerk et al., 2009; Shatilova et al., 2010). Palynological records
102 document a coastal plain vegetation and generally outline warm and humid conditions; however,
103 while Connor et al. (2007) describe a gradual evolution from chestnut-dominated open woods in the
104 Mid-Holocene followed by shrubby marshy landscapes and, caused by groundwater fluctuations due
105 to coastal subsidence and sea-level changes, recent beach-hornbeam conditions, de Klerk et al.
106 (2009) highlight a relatively constant vegetation history with only minor effects of anthropogenic
107 land use change. Other studies investigated the degree of anthropogenic landscape changes, e.g.
108 deforestation (Nikolaishvili et al., 2011), human-induced landslides (Nikolaeva et al., 2014; Vezolli et
109 al., 2014), and soil contamination (Narimanidze and Brückner, 1999; Svanidze et al., 2008). Several
110 geoarchaeological projects have been carried out concerning the environs of archaeological sites –
111 e.g. the anthropologically famous Dmanisi (Messager et al., 2009; 2011); Bronze age settlements on
112 the Bedeni Plateau (Kvavadze et al., 2015) – as well as the analysis of fluvial archives (von
113 Suchodoletz et al., 2015).

114 The Holocene reconnection of the Black Sea with the Mediterranean Sea occurred around ~7400–
115 6400 cal BC (Ryan, 2007; Giosan et al., 2009; Lericolais et al., 2009; in a previous publication a date of
116 ~5500 cal BC had been suggested by Ryan et al., 1997). Several studies were dedicated to the
117 subsequent sea-level rise, for which both a steadily rising (Hiscott et al., 2002; Giosan et al., 2009)
118 and an oscillating (Balabanov, 2007; Chepalyga, 2007; Yanko-Hombach et al., 2007) curve were
119 proposed. Based on his evaluation of ~400 radiocarbon ages from Black Sea coasts, Balabanov (2007)
120 suggested several major regression-transgression cycles for the Holocene period. This interpretation
121 was strongly opposed. Brückner et al. (2010) presented evidence for a gradual sea-level rise; they
122 attributed the assumed regression-transgression wiggles of the Balabanov curve to local
123 neotectonics and a misinterpretation of the data set (see also Kelterbaum et al., 2012). Especially the
124 concept of the so-called Phanagorian Regression, which was postulated for the time span ~800 BC to
125 ~500 BC, i.e. for the time when the Greek settlers had founded many coastal colonies in the Black
126 Sea region, was rejected (Fouache et al., 2012).

127 For the Kuban delta plain (Taman Peninsula, Russia), Brückner et al. (2010) proposed a gradual sea-
128 level rise from 7 to 6 m below present sea level at 5000 BC to ~2 m below sea level at 2000 BC, and a
129 subsequent decelerated rise until today. For the Danube delta region, Giosan et al. (2006) assume a
130 stable (± 1.5 m) sea level over the last 5000 years, based on morphodynamic and palaeogeographic
131 data. As in other study areas, the local sea-level evolution is assumed to have influenced both
132 sedimentation patterns and settling activities in Western Georgia as well.

2.3 Human occupation and archaeological background

134   In the Colchian plain, findings of Lower Palaeolithic (~12000 BC) tools indicate a long occupation
135   history of the Kolkheti lowlands (Fähnrich, 2010). Although the transition from hunter-gatherer
136   communities to a pastoral and agricultural sedentism is supposed to have taken place in the Neolithic
137   (8th to 5th millennium BC) only few of these sites are known to date. Increasing numbers of
138   settlements are know from the Late Chalcolithic and Early Bronze Age, the latter beginning in the first
139   half of the 3rd millennium BC (Lordkipanidze, 1991; Fähnrich 2010). Henceforth, settlement mounds
140   occur in the northern part of the Kolkheti lowlands, which represent typical dwelling places of the
141   Bronze Age society in Western Georgia (Jibladze, 2007; Fähnrich, 2010; Gamkrelidze, 2012). In the
142   first half of the 2nd millennium BC, the Colchian culture was formed, which was accompanied by an
143   advance of technical innovations. It merged into the Kingdom of Colchis with a cultural heyday in the
144   8th century BC (Lordkipanidze, 1991). At the same time, ancient Greek merchants reached the Black
145   Sea area and founded several colonies along the Colchian coast. Early cultural exchange (supposedly
146   originating from Mycenaean time) is likely reflected in the saga of the Argonauts, in which the Colchis
147   is praised for its metal (particularly gold; 'Golden Fleece') works and its prosperous society
148   (Lordkipandize, 1991).

149   The city of Phasis represents the most important ancient Greek colonial foundation in the Colchian
150   territory. It was described by various ancient authors such as Strabo, Pliny the Elder, Pseudo-
151   Scymnos, Pseudo-Scylax and Hippocrates. It was probably situated in the Rioni delta close to Lake
152   Paliastomi (Fig. 1) (Gamkrelidze, 1992, 2012; Lordkipanidze, 2000; Korenjak, 2003). However, while
153   the location of ancient Phasis has been related to ruins of a Roman fort (Montpéreux 1842) which
154   existed until the construction of the former airport of Poti, remnants of the city hitherto have not
155   been discovered, and its exact position remains unclear.

156

157   **3 Methods**

158   *3.1 Field work*

159   Our research is based on eleven sediment cores and two trenches, organized in two transects
160   (transect A: general direction W − E; Figs. 1–3, 5; transect B: general direction N − S; Figs. 1, 3–6).
161   Corings were done with the Cobra TT (Atlas Copco) percussion coring device. The sediment cores
162   reached a maximum depth of 12 m below surface (b.s.), while core diameters were 6 and 5 cm,
163   respectively. The trenches reached 2 m b.s. Field documentation included the description of
164   sediment texture, colour, and the $CaCO_3$ test (with HCl, 10%). Samples were taken from the different
165   sedimentary units, and at regular spacing intervals of 30-50 cm.

166   Three master cores which are representative of the study area were analysed in detail: Core KUL 3
167   (Fig. 2) represents the sand-dominated stratigraphy close to the recent shoreline and is situated in
168   the direct vicinity of trench KUL 2 in the slope of the easternmost beach-foredune ridge facing the
169   swampy hinterland. Cores KUL 7 and PAPO 2 represent the typical depositional sequence of the
170   coastal hinterland. KUL 7 (Fig. 3) is situated in the eastern central part of the swamp, ca. 100 m north
171   of the Tsiva river, an artificial drainage channel which crosses the swampy lowlands from southeast
172   to northwest. Core PAPO 2 (Fig. 4) is located close to the Rioni, some 7 km south of KUL 7. From the
173   three master cores, plant fragments and charcoal remains were taken for radiocarbon dating. The
174   two trenches, KUL 1 and KUL 2 (Fig. 5), were dug on top of the most landward lying beach-foredune
175   ridges just north of the Rioni river mouth.

176 The elevation of cores and trenches was measured by using a Topcon Hiper V DGPS with a spatial
177 resolution of 2 cm. Soviet topographic maps (1:25,000 scale, approx. 1960s; 1:50,000 scale, approx.
178 1970s) served for supplementary information on surface changes over the last decades.

179 *3.2 Geochemical and sedimentological analyses*

180 They were conducted at the Geo-Laboratory of the Institute of Geography, University of Cologne. All
181 samples were dried at 40 °C for 48 h, sieved to <2 mm, and gently pestled by hand for aggregate
182 disintegration. Granulometric analyses were conducted for the fine-grained fraction (<2 mm) of all
183 samples using a Laser Diffraction Particle Size Analyzer (LS 13320 Beckmann Coulter$_{TM}$). Samples
184 were pre-treated with hydrogen peroxide (H_2O_2 15 %) to remove the organic matter, and sodium
185 pyrophosphate ($Na_4O_7P_2$, 46 g/l) was added as a dispersant to avoid coagulation. Each sample was
186 measured three times using the optical Fraunhofer model. Grain-size parameters based on Folk and
187 Ward (1957) were calculated using GRADISTAT software version 8 (Blott and Pye, 2001).

188 Loss on ignition (LOI) was determined by oven-drying of 5 g sample material at 105 °C for 12 h and
189 ignition in a muffle furnace (Carbolite ELF) at 550 °C for 5 h. Although possible uncertainties may
190 result from the combustion of clay minerals and/or carbonates, LOI was used for estimating organic
191 carbon contents in the sediments (Barsch et al., 2000; Heiri et al., 2001).

192 Furthermore, the samples of core KUL 7 were analysed for their C/N ratio using the method
193 described by Meyers and Teranes (2001). Total organic carbon (TOC), total carbon (TC), and nitrogen
194 (N) were determined on duplicate powdered samples using a Vario EL Cube (Elementar
195 Analysensysteme GmbH, Hanau, Germany). Before measurement the material was homogenized and
196 weighed out into tin boats. In a second aliquot of each sample $CaCO_3$ was dissolved with 10 % HCl,
197 before TOC was determined.

198 In addition, selected samples were analysed for their mineral composition by X-ray diffraction (XRD;
199 cores KUL 3 and 7; Fig. 7C) and for their element contents by X-ray fluorescence (XRF; cores PAPO 2,
200 KUL 3 and 7). The material was homogenized using an automatic ball grinder (Retsch MM 400),
201 pressed into 2 mm thick pellets, and measured with a portable XRF analyser (NITON XL3t). Each
202 sample was measured three times in mineral mode for 160 sec to cover all possible filter options
203 with an adequate time span. In case of the elements K, Ca, Fe, Mn errors were below 1 %, in case of
204 S below 4 %. For XRD analysis, the homogenised powder was placed on a PVC slide and measured in
205 a Powder X-Ray Diffractometer (Siemens D 5000) with a fixed focal distance of 0.5 mm. All samples
206 were measured at 5-75° 2Θ in 0.05° steps and 10 sec per degree (Cu-K-alpha radiation source,
207 operated at 40 keV and 40 mA). The data were analysed using the DiffracPlus EVA software package
208 (Bruker AXS, Berlin, Germany).

209 To distinguish the different facies two forms of visualisation were chosen. For the principal
210 component analysis (PCA; Fig. 7A) the PAST software (version 3.1.1) was applied (Hammer et al.,
211 2001). The values were standardized and standard deviation was calculated. Altogether eight
212 components (contents of K, Ca, Fe, Mn, LOI mud and sand, and sorting) were used. The spatial
213 distribution bases on the principle components 1 and 2. The ternary plot (see Brumsack, 1989;
214 Dellwig et al., 2000; Fig. 7B) was evaluated with the Grapher 9 software and predicates on the
215 standardized values of mean grain size, sorting and Ca/K ratio.

216 Finally, six samples of core KUL 12 were analysed for their microfaunal content. Samples were sieved
217 into fractions of >100 μm and 63-100 μm after adding sodium pyrophosphate ($Na_4O_7P_2$, 46 g/l) to
218 avoid coagulation. Species determination was carried out under a light stereo microscope.

219 *3.3 Dating techniques*

220   Thirteen samples (plant remains and charcoal) taken from cores KUL 3, KUL 7 and PAPO 2, were
221   radiocarbon-dated in the CologneAMS Laboratory, University of Cologne (Tab. 1). All ages were
222   calibrated with the Calib 7.1 software (calibration data set: intcal13.14c, following the approach of
223   Reimer et al., 2013). The age-depth-model was calculated with the OxCal version 4.2.4 (Bronk
224   Ramsay and Lee, 2013), restricted to the deposition models for chronological records of Bronk
225   Ramsay (2008).

226   In order to gain chronological information on the generation and last activation of the beach-
227   foredune ridges north of the Rioni river mouth, infrared stimulated luminescence (IRSL) dating was
228   carried out on four samples from trenches KUL 1 and 2. The luminescence samples were taken by
229   pushing opaque steel tubes of 5 cm diameter into the cleaned vertical sections. The tubes were
230   closed with plastic lids to avoid light contamination. For further information on sample preparation
231   see Appendix 1.

232

233   **4 Results**

234   *4.1 The beach-foredune ridges and the adjacent hinterland (KUL 1 – KUL 3)*

235   KUL 1 and 2 were trenched down to a depth of ca. 2 m below surface (b.s.). They comprise of
236   homogeneous coarse to medium sand layers, with a few roots. Only the upper ~30 cm reveal a
237   slightly finer grain size and an increased content of organic matter. Sediment core KUL 3 (Fig. 2) is
238   composed of well-sorted medium sand (mean grain size between 522 and 455 µm) with organic
239   matter of <4 % (6.70-2.82 m b.s.) at its base. Subsequently, the mean grain size gradually decreases,
240   and the organic content varies between 2.82 and 2.46 m b.s. The overlying peaty unit (2.46-1.95 m
241   b.s.) shows grain size values of 254-237 µm and high LOI (15-67 %). Above, relatively homogeneous
242   sand occurs. The uppermost part (0.38–0 m) is characterised by decreasing mean grain size and
243   increased LOI.

244   *4.2 The Kolkheti wetlands – master cores KUL 7 and PAPO 2*

245   *4.2.1 Sediment core KUL 7*

246   The lowermost stratigraphical unit (12-11.68 m b.s.) is composed of grey homogeneous silt which is
247   characterized by high Ca contents, also reflected in the Ca/Fe and Ca/Ka ratios (Fig. 3). The mineral
248   composition of this layer differs from the subsequent layers; calcite was detected only in sample
249   KUL7/44. The overlying stratum (11.68-11.55 m b.s.) consists of an intensely weathered dark-brown
250   peat with high LOI values (up to 37 %) and TOC/N ratios, and low Ca/K and Ca/Fe ratios (Fig. 3). At
251   11.55 m b.s. the peat gradually changes into dark grey clayey silt (mean grain size <10 µm) with few
252   shell fragments and elevated contents of organic matter, particularly between 11.31 and 11.28 m b.s.
253   At 11.00 m b.s., LOI reaches a constant level below 5 %, while Ca remains high and TOC/N decreases.
254   The clayey silt dominates until 7.71 m b.s., although it is intercalated by a dark brown peat at 9.56–
255   9.40 m b.s. (LOI up to 23 %). Subsequently, the sand content increases considerably, and fine sand-
256   dominated layers alternate with silt-dominated layers until 5.27 m b.s. While LOI and TOC/N remain
257   relatively low, Ca/Fe and Ca/K show constantly decreasing values above 6.43 m b.s. In general,
258   elevated Si and Ti values characterise the strata between 9.40 and 5.47, compared to the sediments
259   below.

260   Above, the sedimentary succession is characterised by two distinct peats (5.27 and 2.49 m b.s.),
261   intercalated by a light grey silt layer (3.89–3.59 m b.s.). The peat units are indicated by increased S
262   and TOC/N; they are mainly composed of a clay- and silt-dominated matrix (mean grain size <9 µm)
263   with high LOI values (10-34 %) and only few macroscopic plant remains. Subsequently, the peat layer

264 gradually changes to clayey silt which prevails until 0.45 m b.s. Above the ground water table, the
265 sediments have a reddish brown colour, contain root fragments and show signs of redox
266 characteristics. The uppermost layer is the plough horizon.

267 *4.2.2 Sediment core PAPO 2*

268 From 10 to 9.78 m b.s. sandy silt with low LOI occurs. It is covered by silty sand with large plant
269 remains, high LOI, Ca/K and Ca/Fe, and low Si and Ti values (Figs. 1, 4). Then follows a peat (LOI 41-79
270 %, 8.59-8.03 m b.s.) with very low Ca/K and Ca/Fe ratios. It gradually changes to grey homogeneous
271 (8.03-~6.00 m b.s.) and laminated (~6.00-3.36 m b.s.) clayey silt, which is characterized by low LOI,
272 but high values of Si, Ti, Ca/K and Ca/Fe. Sandy intercalations are found, e.g. between 6.39 and 6.03
273 m b.s. Then peat layers with macroscopic plant remains occur between 3.36 and 2.76 m b.s., and
274 from 2.22 to 1.41 m b.s.; they are characterised by increased organic contents and reduced Ca/K and
275 Ca/Fe ratios. The peats are separated by grey clayey silt (2.76-2.22 m b.s.). Red-brown silt with low
276 LOI occurs in the uppermost ~2 m of the core.

277 *4.3 Additional stratigraphic information along the coring transects A and B*

278 Compared to master cores KUL 7 and PAPO 2, a number of similarities can be found in the
279 sedimentary succession of all other cores along the two coring transects (Figs. 5, 6). All cores are
280 predominantly composed of fine-grained (silt-dominated) and often laminated sedimentary units
281 with macroscopic plant remains. However, depending on the location of each core along the coring
282 transects, differences exist regarding the presence of basal sand layers as well as the position of sand
283 and peat intercalations.

284 In the southern section of coring transects A and B, the lower parts (below 5.5 m b.s.l.) of cores SHAV
285 2 and PAPO 3 (Fig. 5) comprise fine sand-dominated and poorly sorted layers with varying silt
286 contents (mean grain size ~30-200 μm), which is similar to the section between 9.78 m and 8.59 m
287 b.s. in master core PAPO 2. In case of SHAV 2, this unit reaches a thickness of almost 3 m. In contrast,
288 units of (silty) sand are better sorted and more homogeneous, which is expressed by reduced
289 fluctuations of the mean grain size in cores KUL 7, 9, 10 and 11. Finally, medium sand with parts of
290 coarse sand (~250-520 μm) are found in core KUL 3, in sediment profiles of trenches KUL 1 and 2, and
291 in the uppermost 2.5 m of core KUL 12. Here, sand units are well-sorted (in general σ <2),
292 homogeneous and void of macroscopic plant remains.

293 The distribution and thickness of peats reveal a certain pattern as well. Except for KUL 6, all cores
294 contain at least one distinct peat layer. The peats generally occur in the upper parts of the cores,
295 predominantly in a sequence of 2-3 layers, which are separated by thin clay- and silt-dominated
296 strata. In the lower core sections, peats were found in only 7 cores.

297 Six samples of core KUL 12 were analysed for their microfaunal content, with only two of them
298 revealing scattered findings: At 11.74 m b.s. a few indeterminable remains of foraminifera occur; at
299 9.49 m b.s. a sample contains two indeterminable foraminifera specimens.

300 *4.4 Radiocarbon and IRSL dating results*

301 Tab. 1 gives the details of the radiocarbon dating results, including specifications of the dated
302 material, sample numbers and depths. The ^{14}C ages (2 σ) are noted in the figures of the profiles at
303 the position of the sampled depths (see Figs. 2, 3 and 4).

304 The fading-corrected IRSL dating of samples from trench KUL 1 resulted in ages of 1010 ± 134 a (KUL
305 1/2) and 1017 ± 129 a (KUL 1/4). The IRSL ages from trench KUL 2 are only insignificantly younger
306 (KUL 2/1: 1047 ± 134 a; KUL 2/2: 1049 ± 134 a) (Tab. 2).

307

5. Discussion

5.1 Facies determination

Based on the samples taken from the master cores, eight facies were distinguished, which were adopted for the further cores of coring transects A and B. The facies represent depositional environments that differ in terms of grain size, sorting, organic content and elemental composition (Figs. 7A, B, C).

Facies A: shallow marine (?)

Facies A was only found in the basal section of KUL 7 (Fig. 3). From 12 m b.s. to the peat layer at 11.68 m b.s., the sediment shows low S values and the presence of (magnesium-) calcite (Fig. 7C), which is manifested in high Ca contents. Samples from this section contain few undetermined shell fragments and few foraminifera. This facies differs from facies D (alluvial, semi-terrestrial deposits) and facies C (lagoonal mud) in terms of lower Si, Ti and S contents. The slightly increased TOC/N ratio is a possible indicator for both shallow marine and terrestrial conditions (Meyers and Teranes, 2001). The PCA and the standardized ternary plot (Figs. 7A, B) both reveal no distinct differentiation of this sample to facies C and D. Therefore, a lagoonal or semi-terrestrial origin, e.g. related to a standing water body on the evolving delta floodplain, comparable to facies C1 or C2, cannot be excluded. In this case a continuous sea-level rise and drowning of the study area would have provoked the growth of a transgressive peat (KUL 7, 11.68-11.55 m b.s.).

Facies B: sublittoral to littoral

Relatively well-sorted medium to coarse sand between 6.70 and 2.82 m b.s. at KUL 3 points to hydrodynamic conditions with rather high sediment transport capacities, typical of wave-dominated sublittoral to littoral conditions (Dingler, 2005; Aavaard and Hughes, 2013). Sediments from facies B are coarser and better sorted than sediments from facies F (fluvial deposits). Based on Folk and Ward (1957), beach sands are characterized by good sorting and tend to unimodal distributions, even if they are located close to river mouths (Tucker, 1996; Hesp, 2002). In the ternary plot this facies is congruent to facies G due to the coarse grain size, but it is distinguishable from the other remaining facies (Fig. 7B). A slightly better differentiation is revealed by the PCA (Fig. 7A), although two samples from facies G plot similar to the samples of facies B as well.

Facies C: lagoonal

Sediments from facies C have a grey colour and tend to be indicated by low contents of organic matter. The lagoonal sediments differ from the semi-terrestrial and shallow marine deposits in terms of higher Ca/Fe and Ca/K ratios and low Si and Ti contents (Figs. 3, 4, 7B). Furthermore, the low TOC/N ratios point to a predominance of algae and low terrestrial input of vascular plant material (Meyers and Teranes, 2001) which supports the differentiation between those two facies.

Nevertheless, several samples from the lagoonal strata reveal a coarser grain size. Although strong storms are known from the Black Sea, neither the absence of microfauna, nor geochemical or sedimentological features typically related to storm deposition (e.g. lamination, fining upward sequences) were found. Therefore a deposition by wave events is unlikely (Morton et al., 2007). However, geochemical parameters are consistent with still water conditions as well. Therefore, facies C can be divided into two subunits: Facies C-1, which consists of the predominating fine-grained mud of the central lagoonal areas, and Facies C-2, which is formed by the coarser deposits of the lagoon areas close to the river mouths. This explains the similar composition between some lagoonal and

350 alluvial samples in the PCA (Fig. 7A). The standardized ternary plot reveals a well-defined distinction
351 of all other facies, except for two fluvial samples which may be explained by the fluvial origin of both
352 facies C-2 and facies F (Fig. 7B).

Facies D: alluvial (overbank deposits)

354 As typical for alluvial sediments, facies D is characterised by silty to clayey deposits. The high TOC/N
355 ratio is an effect of the increased content of vascular or even cellulosic plant material of terrestrial
356 origin (Meyers and Teranes, 2001). Extensive inundation of low-lying delta areas and oxbow lakes
357 during flood events typically induce the accumulation of fine sediments from slack waters as
358 overbank deposits (overbank fines), when the stream overtops its banks (Blair and McPherson,
359 1994). In any case, major floods of the lower sections of the rivers Rioni and Khobistsqali can also
360 provoke crevasse-splays by breaching of levees, most commonly leading to sand deposition on the
361 floodplain (e.g., North and Davidson, 2012). However, crevasse-splay deposition typically goes along
362 with basal erosional unconformities, and no such erosional surface was found in the presented cores.
363 Core sections with varying fine sand contents and Ca/Fe and Ca/K values thus reflect the fluctuating
364 hydrodynamic conditions on the delta plain which are predominantly characterised by the deposition
365 of overbank fines (Boggs, 2006). These coarser layers have similar granulometric characteristics
366 compared to facies C-2, but can be distinguished by differences in the geochemistry. Due to high LOI,
367 many samples resemble those of facies E (Figs. 7A, B).

Facies E: semi-terrestrial (peat)

369 Facies E is represented by peat with numerous macroscopic plant remains and considerably elevated
370 LOI and K values. Preservation of plants points to anoxic conditions (Turney et al. 2005). High TOC/N
371 ratios are typical of peat bogs (Joosten et al., 2003; Haberl et al., 2006) and indicate a dominance of
372 cellulosic plants (Meyers and Teranes, 2001). The most remarkable difference is revealed in elevated
373 LOI and S values. The connection with the surrounding strata is reflected in the scattered distribution
374 in the PCA and ternary plot (Fig. 7A, B), e.g. adjacent to facies B, C-1 and C-2.

Facies F: fluvial

376 A fluvial origin is assumed for some sections of cores PAPO 2, PAPO 3 and SHAV 2 and refers
377 primarily to sediment input by distributary channels of the two rivers in the delta. These sediments
378 show coarser grain size, moderate to poor sorting (Sun et al., 2002), and are void of macro- and
379 microfauna. Elevated Fe and K values point to increased input of terrestrial material (Arz et al., 1998;
380 Meyers and Teranes, 2001; Kujau et al., 2010). The varying grain size and sporadically occurring
381 fining-upward sequences (e.g. PAPO 2) can for instance be explained by river avulsion and temporary
382 lateral channel migration, deposition during larger floods in areas close to distributary channels,
383 and/or – in contrast to facies D – crevasse splay deposition (Boggs, 2006). There are certain
384 similarities with facies C-2 (Fig. 7B), but facies F reveals coarser mean grain size and elevated LOI, Si
385 and Ti values. A better separation is revealed in Fig. 7A, where the plots of facies F and C-2 are
386 disjoint.

Facies G: aeolian

388 Facies G is restricted to sediment profiles of trenches KUL 1 and 2, and the upper part of KUL 3. The
389 elevation of these sediments with respect to sea level (2.50-3.00 m a.s.l.), the high degree of sorting
390 and the dominance of medium sand (mean grain size 310–410 µm) suggest an aeolian origin. The
391 location on the landward slope of the shore-parallel foredune ridges confirms this assumption (Hesp,
392 2002). The high degree of similarity to littoral sediments (facies B; Figs. 7A, B) can be explained by

393 the shared source area and the short transport distance. This is reflected by slightly increased fine
394 sand and reduced coarse sand contents (Fig. 2).

395 *Facies H: anthropogenic*

396 This facies reveals the same characteristics as facies B (sublittoral to littoral environments). The
397 sediment was intentionally deposited at the site of KUL 12 in order to build a causeway.

398 *5.2 Chronostratigraphic interpretation of sediment cores*

399 *5.2.1 Trenches KUL 1 and 2, core KUL 3*

400 In KUL 3 the coastal evolution north of the present Rioni mouth is documented for the last 3500
401 years (Fig. 2). Due to the circulation pattern of the offshore currents (Korotaev et al., 2003), the Rioni
402 is the main sediment source for the littoral deposits along the graded shoreline. Relatively well-
403 sorted medium to coarse sands between 6.70 and 2.82 m b.s. point to relatively high transport
404 capacities. Sublittoral to littoral conditions (facies B) at site KUL 3 established before ~1500 cal BC
405 and ceased some time before ~350-50 cal BC, when decreasing mean grain size (3.25 m and 2.82 m
406 b.s.; facies B and C) and a subsequent peat layer (2.82-1.95 m b.s.; facies E) indicate a back-barrier
407 milieu with beach aggradation and/or delta progradation. These conditions continued at least until
408 ~500 cal AD. Elevated S values and well-preserved plant remains suggest an anoxic milieu in the back-
409 barrier swamp.

410 After ~500 AD, well-sorted medium sand was accumulated at site KUL 3, which is similar to the
411 sediment in the upper 2 m of the adjacent foredune ridges (KUL 1 and 2, facies G) and therefore of
412 aeolian origin. Based on the IRSL ages taken from the trenches, sand deposition on top of the
413 foredunes was still active or was reactivated around AD ~830 – 1150 AD (1049 ± 134 a and 1010 ±
414 134 a, Tab. 2), before the eastern part of the ridges was finally stabilized. This is roughly in
415 accordance with the uppermost radiocarbon age of KUL 3: ~500 cal AD is a *terminus post quem* for
416 the deposition of the well-sorted medium sand.

417 However, several uncertainties may be related to the IRSL ages. Due to the low temperature of the
418 preheat plateau not all narrow electron traps may have been emptied, which could lead to age
419 overestimation (Auclair et al., 2003). Furthermore, the feldspar aliquots show a significant scatter of
420 equivalent doses. Since the samples were taken from the upper section of the foredune ridges and
421 likely experienced aeolian transport, sediments are assumed to be rather well-bleached. Therefore,
422 post-depositional mixing may be a reason for the large scatter. The reforestation of the area with
423 pine trees during the 20th century (De Klerk et al., 2009) and resulting root penetration could be a
424 possible explanation. Thus, the IRSL ages should be treated as maximum ages for the last period of
425 foredune activity.

426 *5.2.2 Master cores KUL 7 and PAPO 2*

427 The development of the central part of the swampy hinterland between the rivers Khobistsqali and
428 the Rioni is documented by master cores KUL 7 (Fig. 3) and PAPO 2 (Fig. 4). For the lowermost section
429 of core KUL 7 (12-11.68 m b.s.), two different scenarios may be considered: assuming a shallow
430 marine environment of the basal unit, the following peat documents the establishment of semi-
431 terrestrial conditions, e.g. in the form of a floodplain swamp (facies D), pointing to a westward shift
432 of the coastal environments. Assuming that the basal unit of KUL 7 represents lagoonal to semi-
433 terrestrial conditions, a western position of the coastline must be assumed, and back-barrier
434 conditions existed at KUL 7 already before ~6000 cal BC.

435 However, as documented by the peat layer, back-barrier conditions dominate the entire study area,
436 except for site KUL 3, at least since 5979-5740 cal BC (Tab. 1). Subsequently, peat growth could not
437 keep pace with the, at that time, still considerably rising sea level (Giosan et al., 2006; Brückner et al.,
438 2010). The establishment of open lagoonal conditions (facies C) is inferred from the following ~4 m
439 thick unit of clayey silt, although the mineralogical composition (i.e. absence of calcite) of these
440 sediments and the absence of micro- and macrofossils point to conditions inappropriate for the
441 development of faunal assemblages. Depositional conditions remained stable for the following
442 centuries, and, despite the development of a thin intercalated peat, were most likely characterised
443 by a lagoonal back-barrier situation. While no microfauna was found to determine salinity conditions,
444 the geochemical parameters are the most reliable indicators for this facies determination. High
445 depositional rates are inferred from the ^{14}C age of plant fragments taken from the uppermost part of
446 the silt layer: with an age between 5529 and 5346 cal BC (Tab. 1) it is only slightly younger than the
447 sample taken from the peat below.

448 In the following units we assume an increasing significance of alluvial processes, e.g. overbank
449 deposition during major floods of the lower sections of the rivers Rioni and Khobistsqali. The
450 increased alluvial deposition marks the transition to subsequently evolving semi-terrestrial
451 conditions (facies D), which were likely related to the gradual evolution of the Rioni-Khobistsqali
452 floodplain, i.e. the Kolkheti lowlands, established around 3600 cal BC. Subsequently, the stratigraphy
453 is dominated by thick peat units (facies E). Peat formation at KUL 7 dominated at least until ~900 cal
454 AD; sporadic sand input on the floodplain is inferred from the sand fraction in the peat layers.

455 Additionally, different rates of peat accretion may be inferred from the radiocarbon ages. While 1.50
456 m of peat accumulated between 3695-3521 cal BC and 3348-3025 cal BC (5.27-3.89 m b.s.) (Tab. 1),
457 reduced rates of peat formation must be assumed for the section between 3.59 and 2.49 m b.s.,
458 where 1.10 m of peat formed during a period of 4000 years (3348-3025 cal BC to 655-861 cal AD). If it
459 is not a dating error, this might be explained by different rates of groundwater level changes.
460 However, the development of paralic peats is dependent on the local sea-level evolution (Fouache et
461 al., 2012), and a sea-level rise of only 1.1 m in ~4000 years is problematic. Thus, possible
462 uncertainties have to be considered, e.g. dating of different material, a possible hiatus within the
463 sequence and floating peat (Haberl et al., 2006; Vött et al., 2007). Finally, the peat is covered by fine-
464 grained alluvial deposits and 0.45 m of anthropogenic infill.

465 The stratigraphy recorded in core PAPO 2 reflects a similar evolution. The stratigraphy starts with
466 lagoonal deposits (facies C). The following sandy stratum of fluvial origin (facies F) was deposited
467 between 5297-5044 cal BC and 3517-3118 cal BC. When the riverine input declined, peat (facies D)
468 formed, indicating semi-terrestrial conditions and the establishment of a lagoonal environment, still
469 affected by episodic flooding. The end of input of coarser-grained sediments may reflect an avulsion
470 of the Rioni. After 1618-1441 cal BC, peat growth indicates the end of the lagoonal conditions (3.36-
471 2.76 m b.s.). Related to the extending delta floodplain of the Rioni, alluvial sediments (facies E) had
472 been deposited at the site at least since 82-322 cal AD.

473 *5.3 The evolution of the Kolkheti lowlands*

474 The evolution of the Kolkheti lowlands can be deduced from coring transects A and B (Figs. 5, 6). The
475 facies of the three master cores were adopted to the cores along the coring transects. In addition,
476 the cores were classified into four groups based on their site-specific sedimentary succession.

477   <u>Group 1</u> consists of core KUL 3 and trenches KUL 1 and 2. It is dominated by sublittoral to littoral (KUL
478   3) and aeolian sands (upper part of KUL 3, KUL 1 and 2). Similar deposits are absent in the other parts
479   of the coring transects.

480   Cores SHAV 2, PAPO 2 and PAPO 3 comprise <u>group 2</u>, which is restricted to the southern part of the
481   research area. In these cores, the lowermost sections (below ~6 m b.s.l.) show lagoonal
482   environments (facies C), which were episodically affected by sediment input from the Rioni river
483   (facies F). Especially in the case of SHAV 2, a 3 m thick layer of fluvial sand occurs, which can be
484   related to similar sand intercalations at PAPO 2 and PAPO 3. Due to ^{14}C-dated plant material in PAPO
485   2 (Tab. 1, sample PAPO2/31), the distinct sand layers was accumulated after ~5300 cal BC. It is
486   assumed that by then the course of the Rioni was located at a similar position as today. When
487   lagoonal conditions (facies C) were established in the area after ~3000 BC, major sand intercalations
488   are absent. Instead, intercalated peats indicate recurrent siltation and shoreline shift of a shallow
489   lagoon. In the southern part of transect B, the transformation from lagoonal (facies C) to semi-
490   terrestrial and later terrestrial depositional environments (facies D and E) occurred after ~1500 BC.

491   Further north, cores KUL 6, 9 and 12 (Fig. 1) comprise <u>group 3</u>. The cores represent the central part of
492   the Kulevi swamps between the rivers Rioni and Khobistsqali. They almost entirely consist of clayey
493   silt, which is interpreted to represent lagoonal (facies C) and, subsequently, alluvial strata (facies E).
494   Alluvial and fluvial deposition was reduced wherefore intercalations of coarser material are missing.
495   It seems that these core sites had been located in the centre of the former lagoon. Only a thin sand
496   layer suggests flood-related deposition at KUL 9. The sand layer in the upper 2.5 m of KUL 12 is
497   artificial infill (facies G; Figs. 5, 7) since the coring site is located in close proximity to the road.

498   <u>Group 4</u> comprises cores KUL 5, 7, 10 and 11 and exhibits parallels with group 2. KUL 7 is the deepest
499   core and reaches ~10.50 m b.s.l. It is the only core where shallow marine conditions (facies A) are
500   inferred at the base. However, the subsequent strata are dominated by fine-grained sediments of the
501   lagoonal facies (facies C-1), with sand intercalations indicating episodic alluvial sediment input (facies
502   C-2) particularly at sites KUL 7 and 10 which are located close the river mouth. While peat layers are
503   almost absent in the lower core sections, thick peat units are present in the upper part of the cores
504   from this group. We have not yet resolved whether alluvial deposition is related to the Rioni or the
505   Khobistsqali.

506   In summary, the presented findings allow to sketch a general scenario for the evolution of the
507   Kolkheti lowlands over the last eight millennia.

508   Around 6000 cal BC, a shallow marine environment may be inferred from the base of two cores (KUL
509   7 and 12); these sediments show slightly different characteristics compared to all other strata.
510   However, samples from this unit cannot be unambiguously separated from the lagoonal and semi-
511   terrestrial samples based on the sedimentological and geochemical characteristics (Fig. 7). It thus
512   remains open whether the Holocene transgression is reflected in the form of shallow marine
513   deposits, or in the form of landward migrating back-barrier conditions.

514   Around 5900 cal BC, the coastline was located west of coring transect B, most likely between KUL 12
515   und 3. In all cores, a thick sequence of fine-grained sediments points to standing water bodies and
516   quiescent depositional conditions, evidencing the existence of a coastal barrier. As suggested by peat
517   intercalations and input from the rivers, recurrent siltation and re-establishing open water bodies
518   characterised this back-barrier environment.

519   Since ~3500 cal BC (at KUL 7), the delta plain is assumed to have reached the area of the coring
520   transects. The final transition from lagoonal to semi-terrestrial environments is represented by the

521 development of peat. In other parts of the lowlands, this change took place significantly later, e.g.
522 ~1500 cal BC at site PAPO 2. Further west, at KUL 3, the siltation was connected to the subsequent
523 westward shift in the shoreline. North of the present Rioni mouth it took place sometime before
524 ~350 cal BC.

525 This scenario is in agreement with the results presented by Connor et al. (2007) and de Klerk et al.
526 (2009), who suggested an interaction of sea-level rise and fluvial infill causing fluctuating
527 groundwater conditions in peat bogs at the site of Ispani. Furthermore, the establishment of a
528 floodplain with peat bogs is indirectly corroborated by historical sources from Antiquity: although
529 uncertainties remain regarding the locations and landscape descriptions given by Hippocrates (c. 460
530 – c. 370 BC) and Strabo (64/63 BC – c. AD 24), both describe the surroundings of Phasis and the Rioni
531 as marshy forests and swamps (cf. Gamkrelidze, 2012).

532 *5.4 Implications for the local relative sea-level evolution*

533 Although the reconstruction of a relative sea-level curve for the research area remains challenging
534 due to the lack of unambiguous sea-level index points, the relationship of each dated sample to the
535 respective past sea level allows for an approximation of the relative sea-level position for each data
536 point (Fig. 8). Based on the facies succession inferred for the cores, the peat in the lower part of core
537 KUL 7 (sample KUL 7/44) is assumed to represent a paralic peat layer. Since paralic peats are directly
538 related to the back-barrier groundwater table, and, thus, to the local relative sea level (Pirazzoli,
539 1991; Vött, 2007), we defined a vertical error range of ~±0.5 m (Pirazzoli, 1996; Behre, 2003) for this
540 and for similar (KUL 3/9, KUL 3/10; back-barrier position) samples. The local relative sea level thus
541 was ca. -9 m around 6000 BC.

542 The three ages of KUL7/37, PAPO2/31 and PAPO2/28 derive from plant remains taken from lagoonal
543 mud (facies C). Since the material was deposited below the former local sea level, these samples
544 represent minimum values for the sea level position at ~5450 BC (-8 m), ~5200 BC (-7.4 m) and ~3300
545 BC (-6.3 m). Rather shallow lagoonal conditions and a deposition close to the lagoonal water table is
546 most likely, since subsequently peat evolved (KUL7/37, PAPO2/31).

547 In contrast, samples taken from an alluvial facies (facies D), i.e. related to the deposition of overbank
548 fines, derive from elevations above the former local sea level. At present, the dry flood plain with
549 episodic overbank deposition shows maximum elevations of ~3 m a.s.l. Therefore, samples KUL 7/20,
550 KUL 7/13, PAPO2/11, PAPO 2/9 and KUL 7/7 indicate maximum values for the respective sea level
551 position. However, while KUL 7/20, KUL 7/13, PAPO2/11 give valuable information about the
552 maximum sea-level position for the time span between ~3500 and ~1500 BC, the latter two samples
553 (PAPO 2/9, KUL 7/7) may be neglected since samples KUL3/10 and KUL 3/9 (paralic peat in back-
554 barrier position) represent more reliable sea-level index points for this time period, with a rather
555 narrow error range. For KUL3/14, representing sublittoral to littoral facies (facies B), an error range
556 of several meters must be assigned as well. Finally, the peat sample PAPO2/6 derives from a core
557 section dominated by alluvial sediments, indicating an oxbow peat, which only gives a maximum
558 estimate for the position of the local sea level.

559 Altogether, our data points (Fig. 8) indicate an estimated relative sea level at ~-9 m ~6000 cal BC, and
560 a subsequent sea-level rise until 2000-1500 cal BC, when a sea level of ~3-2 m below present level
561 was reached. Afterwards sea-level rise decelerated until it reached today's level.

562 Compared with the local sea-level curves for the Taman Peninsula in SW Russia (sites Semebratnee
563 [Brückner et al., 2010], Golobitskaya [Kelterbaum et al., 2011], and Dschiginka [Fouache et al., 2012])
564 (Fig. 8), our curve shows a rather similar trend, in particular when compared to the RSL curves of the

565 former two sites which point to a relatively slow sea-level rise throughout the last ~4000 years as
566 well as to other RSLs of the Eastern Mediterranean (Vött et al., 2007; Brückner et al., 2010).
567 However, the data presented here point to an even slower local relative sea-level rise in the Rioni
568 delta. The differences can be explained by the different tectonic settings of the central Georgian
569 coast and the Taman Peninsula. Sites on Taman Peninsula are related to subsidence (i.e, related to
570 tectonic subsidence and/or compaction) according to Fouache et al. (2012). This may also be the case
571 for the study area as a back-arc marginal extensional basin between the Caucasus ridges (Yılmaz et
572 al., 2013), though at a more moderate pace. Based on the findings presented here, the high
573 subsidence rates suggested by Gamkrelidze (1998) for the study area of 2-6 mm/a may be
574 questioned.

575 Finally, as already suggested by the studies of Giosan et al. (2006), Brückner et al. (2010), and
576 Kelterbaum et al. (2011, 2012), no remarkable wiggles of the curve, representing regression-
577 transgression cycles, can be inferred, which further challenges the theory of the presumed
578 Phanagorian Regression (e.g. Balabanov, 2007).

579

580 **6. Conclusion**

581 Based on a total of eleven vibracores and two sediment trenches, this study presents, for the first
582 time, detailed information on the chronostratigraphy of the Kolkheti lowlands (Colchis) in Georgia.
583 Using a combination of sedimentological and geochemical parameters, we identified different facies,
584 the succession of which reflects the palaeoenvironmental evolution of these coastal lowlands.
585 Radiocarbon and IRSL ages show that the sedimentary record presented here covers the last ~8000
586 years.

587 During this period, the area of research was subject to considerable landscape changes. While a
588 shallow marine origin for the very basal part of cores KUL 7 and KUL 12 remains ambiguous, the rapid
589 establishment of lagoonal conditions, sometime before ~5000 cal BC, can be demonstrated based on
590 sedimentological indicators. The lagoon was separated from the Black Sea by a barrier. Its
591 sedimentation was influenced by the discharge of the rivers Rioni and Khobistsqali. Subsequently,
592 alluviation transformed the lagoon into a delta plain with wetland-dominated sections, characterised
593 by peat bogs, and sections with fine-grained alluvium. Depending on the location within the delta
594 area, this transformation took place between 3500 and 1500 cal BC. Today, a substantial part of the
595 area is still occupied by wetlands with peat bogs and forests, while other parts remain open waters
596 or have been transformed into farmland.

597 As for the sea-level evolution, no evidence for significant oscillations of the local relative sea level
598 was found, in particular no regression-transgression cycles as suggested by Balabanov (2007).
599 Instead, our results point to a progressively and moderately rising sea level, with decelerated speed
600 since 3000-2000 BC. This evolution resembles sea-level curves inferred for the Taman Peninsula
601 (Brückner et al., 2010; Kelterbaum et al., 2011; Fouache et al., 2012). Differences between the curves
602 may potentially reflect different tectonic settings.

603 Against the background of the early establishment of a permanently changing delta plain, it remains
604 difficult to draw conclusions about the possible location for the, as yet, undiscovered location of the
605 ancient city of Phasis. Due to the persistence of extended peat bogs and open water bodies, a
606 location close to the Rioni, for example on a palaeo-levee of the river, seems likely. However, ancient
607 reports locate Phasis south of the Rioni (Gamkrelidze, 1992, 2012; Lordkipanidze, 2000; Korenjak,

608 2003). Further research, including geophysical surveying techniques to detect subsurface
609 palaeochannels of the Rioni, is needed to find the lost city of Phasis.

610 .Based on this study, future research will reveal more detail on the sea-level evolution of Georgia's
611 coast, thereby making an important contribution to the ongoing debate on the sea-level evolution of
612 the Black Sea throughout the Holocene. Plus, the peat bogs to the north of the Rioni and the Lake
613 Paliastomi are promising archives for palynological research and will add new information on
614 Georgia's Holocene vegetation history

615

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References

628 Adamia, S., Zakariadze, G., Chkhotua, T., Sadradze, N., Tsereteli, N., Chabukiani, A., Gventsadze, A.,
629 2011. Geology of the Caucasus: A Review. Turkish Journal of Earth Sciences 20, 489-544.

630 Adamiec, G., Aitken, M., 1998. Dose rate conversion factors: update. Ancient TL 16, 37-50.

631 Auclair, M., Lamothe, M., Huot, S., 2003. Measurement of anomalous fading for feldspar IRSL using
632 SAR. Radiation Measurements 37, 487-492.

633 Arz, H.W., Pätzold, J., Wefer, G., 1998. Correlated millennial-scale changes in surface hydrography
634 and terrigenous sediment yield inferred from last-glacial marine deposits off Brazil. Quaternary
635 Research 50, 157-166.

636 Balabanov, I.P., 2007. Holocene sea-level changes of the Black Sea. In: Yanko-Hombach, V., Gilbert,
637 A.S., Panin, N., Dolukhanov, P.M. (eds.), The Black Sea Flood Question: Changes in Coastline, Climate,
638 and Human Settlement. Springer, Dordrecht, pp.711-730.

639 Balescu, S., Lamothe, M., Auclair, M., Shilts, W.W., 2001. IRSL dating of Middle Pleistocene
640 interglacial sediments form southern Quebec (Canada) using multiple and single grain aliquots.
641 Quaternary Science Reviews 20, 821-824.

642 Barsch, H., Billwitz, K., Bork, H., 2000. Arbeitsmethoden in der Physiogeographie und Geoökologie.
643 Klett-Perthes, Gotha.

644 Bazhenov, M.L., Burtman, V.S., 2002. Eocene paleomagnetism of the Caucasus (southwest Georgia):
645 oroclinal bedding in the Arabian syntaxis. Tectonophysics 344, 247-259.

646 Behre, K.-E., 2003. Eine neue Meeresspiegelkurve für die südliche Nordsee. Transgressionen und
647 Regressionen in den letzten 10.000 Jahren. Probleme der Küstenforschung im südlichen
648 Nordseegebiet 28, 9-63.

649 Bell, W.T., 1979. Attenuation factors for the absorbed radiation dose in quartz inclusions for
650 thermoluminescence dating. Ancient TL 8, 2-13.

651 Blair, T.C., McPherson, J.G., 1994. Alluvial fans and their natural distinction from rivers based on
652 morphology, hydraulic processes, sedimentary processes, and facies assemblages. Journal of
653 Sedimentary Research A64, 450-489.

654 Blott, S.J., Pye, K., 2001. Technical communication, GRADISTAT: A grain size distribution and statistics
655 package for the analysis of unconsolidated sediments. Earth Surface Processes and Landforms 26,
656 1237-1248.

657 Boggs, S., 2006. Principles of Sedimentology and Stratigraphy. Pearson Prentice Hall, New Jersey.

658 Bøtter-Jensen, L., McKeever, S.W.S., Wintle, A.G., 2003. Optically Stimulated Luminescence
659 Dosimetry. Elsevier, Amsterdam.

660 Bronk Ramsay C., 2008. Deposition models for chronological records. Quaternary Science Reviews 27,
661 42-60.

662 Bronk Ramsay, C., Lee, S., 2013. Recent and planned developments of the program Oxcal.
663 Radiocarbon 55, 720-730.

664 Brückner, H., Kelterbaum, D., Marunchak, O., Porotov, A., Vött, A., 2010. The Holocene sea level story
665 since 7500 BP – lessons from the Eastern Mediterranean, the Black and the Azov Seas. Quaternary
666 International 225, 160-179.

667 Brumsack, H.-J., 1989. Geochemistry of recent TOC-rich sediments from the Gulf of California and the
668 Black Sea. Geol. Rundschau 78, 851-882.

669 Carozza, J.-M., Micu, C., Mihail, F., Carozza, L., 2012. Landscape change and archaeological
670 settlements in the lower Danube valley and delta from early Neolithic to Chalcolithic time: A review.
671 Quaternary International 261, 21-31.

672 Chepalyga, A.L., 2007. The late glacial great flood in the Ponto-Caspian basin. In: Yanko-Hombach, V.,
673 Gilbert, A.S., Panin, N., Dolukhanov, P.M. (eds.), The Black Sea Flood Question. Springer, Dordrecht,
674 pp. 119-148.

675 Connor, S., Thomas, I., Kvavadze, E., 2007. A 5600-yr history of changing vegetation, sea levels and
676 human impacts from the Black Sea coast of Georgia. The Holocene 17, 25-37.

677 Cordova, C., Gerasimenko, N., Lehman, P., Kliukin, A., 2011. Late Pleistocene and Holocene
678 paleoenvironments of Crimea: Pollen, soils, geomorphology, and geoarchaeology. GSA Special Papers
679 473, 133-164.

680 Degens, E.T., Ross, D.A., 1972. Chronology of the Black Sea over the last 25000 yrs. Chemical Geology
681 10, 1-16.

682 Dietze, E., Hartmann, K., Diekmann B., IJmker, J., Lehmkuhl, F., Opitz, S., Wünnemann, B., Borchers,
683 A., 2012. An end-member algorithm for deciphering modern detrital processes from lake sediments
684 of Lake Donggi Cona, NE Tibetan Plateau, China. Sedimentary Geology 243-244, 169-180.

685 Dellwig, O., Hinrichs, J., Hild, H., Brumsack, H.-J., 2000. Changing sedimentation in tidal flat sediments
686 of the southern North Sea from the Holocene to present: a geochemical approach. Journal of Sea
687 Research 44, 195-208.

688 Dingler, JR. 2005. Beach processes. In: Schwartz M.L. (eds.) Encyclopedia of Coastal Sciences.
689 Springer, Dordrecht. 161-168.

690 Durcan, J.A., King G.E., Duller G.A.T., 2015. DRAC: Dose rate and age calculator for trapped charge
691 dating. Quaternary Geochronology 28, 54-61.

692 Fähnrich, H., 2010. Geschichte Georgiens. Handbuch der Orientalistik, Bd. 21, Koninklijke Brill NV,
693 Leiden.

Formatiert: Deutsch (Deutschland)

694 Folk, R.L., Ward, W.C., 1957. Brazos river bar: A study in the significance of grain size parameters,
695 Journal of Sedimentary Petrology 27, 3-26.

696 Fouache, E., Kelterbaum, D., Brückner, H., Lericolais, G., Porotov, A., Dikarev, V., 2012. The Late
697 Holocene evolution of the Black Sea – a critical view on the so-called Phanagorian regression.
698 Quaternary International 266, 162-174.

699 Galbraith, R., Roberts, R., Laslett, G., Yoshida, H., Olley, J., 1999. Optical dating of single and multiple
700 grains of quartz from Jinmium Rock Shelter, Northern Australia: Part I, Experimental design and
701 statistical models. Archaeometry 41, 339-364.

702 Gamkrelidze, G., 1992. Hydroarchaeology in the Georgian Republic (the Colchian Littoral). The
703 International Journal of Nautical Archaeology 21, 101-109.

704 Gamkrelidze, G., 1998. Stress Vector Orientations and Movement of the Earth's Crust of Territory of
705 Georgia and the Neotectonic Stage. Bulletin of the Georgian Academy of Science 158, 284-287.

706 Gamkrelidze, G., 2012. Researches in the Iberia-Colchology (History and archaeology of ancient
707 Georgia). Georgian National Museum, Tbilisi.

708 Giosan, L., Donnelly, J.P., Filip, F., Ovejanu, I., Vespremeanu-Stroe, A., Vespremeanu, E., Duller,
709 G.A.T., 2006. Young Danube delta documents stable Black Sea level since the middle Holocene:
710 Morphodynamic, paleogeographic, and archaeological implications. Geology 34, 757-760.

711 Giosan, L., Filip, F., Constantinescu, S., 2009. Was the Black Sea catastrophically flooded in the early
712 Holocene? Quaternary Science Reviews 28, 1-6.

713 Haberl, A., Kahrmann, M., Mutadze, I., Krebs, M., Joosten, H., 2006. The imnati Mire in the Kolkheti
714 Lowland in Georgia. Peatland International 1, 35-38.

715 Hammer, Ø., Harper, D.A.T., Ryan, P.D., 2001. PAST: Paleontological Statistics Software Package for
716 Education and Data Analysis. Palaeontologia Electronica 4, 9-11.

717 Heiri, O., Lotter, A., Lemcke, G., 2001. Loss on ignition as a method for estimating organic and
718 carbonate content in sediments: reproducibility and comparability of results. Journal of
719 Paleolimnology 25, 101-110.

720 Hesp, P., 2002. Foredunes and blowouts: initiation, geomorphology and dynamics. Geomorphology
721 48, 245-268.

722 Hiscott, R.N., Aksu, A.E., Yaşar, D., Kaminski, M.A., Mudie, P.J., Kostylev, V.E., MacDonald, J.C., Isler,
723 F.I., Lord, A.R., 2002. Deltas south of the Bosphorus Strait record persistent Black Sea outflow to the
724 Marmara Sea since 10 ka. Marine Geology 190, 95-118.

725 Huntley, D.J., Baril, M.R., 1997. The K content of the K-feldspars being measured in optical dating or
726 in thermoluminescence dating. Ancient TL 15, 11-13.

727 Huntley, D.J., Lamothe, M., 2001. Ubiquity of anomalous fading in K-feldspars and the measurement
728 and correction for it in optical dating. Canadian Journal of Earth Sciences 38, 1093-1106.

729 Janelidze, C.P., 1980. Paleogeography of Georgia in the Holocene. Metsniereba (in Russian).

730 Jibladze, L., 2007. II-I millennium BC Settlements of Kolkheti lowland, Tbilisi. (in Georgian).

731 Joosten, H., Kaffke, A., Matchutadze, I., 2003. The mires of the Kolkheti lowlands (Georgia). IMCG
732 Newsletter 2003/3, 19-23.

733 Kelterbaum, D., Brückner, H., Porotov, A., Schlotzhauer, U., Zhuravlev, D., 2011. Geoarchaeology of
734 Taman Peninsula (SW Russia) – the example of the ancient Greek settlement of Golubitskaya 2. Die
735 Erde 142, 235-258.

736 Kelterbaum, D., Brückner, H., Dikarev, V., Gerhard, S., Pint, A., Porotov, A., Zin'ko, V., 2012.
737 Palaeogeographic changes at Lake Chokrak on the Kerch Peninsula, Ukraine, during the Mid- and
738 Late-Holocene. Geoarchaeology 27, 206-219.

739 Klerk, P. de, Haberl, A., Kaffke, A., Krebs, M., Matchutadze, I., Minke, M., Schulz, J., Joosten, H., 2009.
740 Vegetation history and environmental development since ca 6000 cal yr BP in and around Ispani 2
741 (Kolkheti lowlands, Georgia). Quaternary Science Reviews 28, 890-910.

742 Korenjak, M., 2003. Die Welt-Rundreise eines anonymen griechischen Autors ("Pseudo-Skymnos").
743 Georg Olms Verlag, Hildesheim.

744 Korotaev, G., Oguz, T., Nikiforov, A., Koblinksy, C., 2003. Seasonal, internannual, and mesoscale
745 variability of the Black Sea upper layer circulation derived from altimeter data. Journal of Geophysical
746 Research 108, 1-15.

747 Kujau, A., Nürnberg, D., Zielhofer, C., Bahr, A., Röhl, U., 2010. Mississippi River discharge over the last
748 ~ 560,000 years—indications from X-ray fluorescence core-scanning. Palaeogeography
749 Palaeoclimatology Palaeoecology 298, 311–318.

750 Kvavadze, E., Sagona, A., Martkoplishvili, I., Chichinadze, M., Jalabadze, M., Koridz, I., 2015. The
751 hidden side of ritual: New palynological data from Early Bronze Age Georgia, the Southern Caucasus.
752 Journal of Archaeological Science: Reports 2, 235-245.

753 Lamothe, M., Auclair, M., 1999. A solution to anomalous fading and age shortfalls in optical dating of
754 feldspar minerals. Earth and Planetary Science Letters 171, 319-323.

755 Lericolais, G., Bulois, C., Gillet, H., Guichard, F., 2009. High frequency sea-level fluctuations recorded
756 in the Black Sea since the LGM. Global and Planetary Change 66, 65-75.

757 Lordkipanidze, O., 1991. Archäologie in Georgien: Von der Altsteinzeit zum Mittelalter. VCH
758 Verlagsgesellschaft mbH, Weinheim.

759 Lordkipanidze, O., 2000. Phasis: The river and city in Colchis. Geographica Historica. Steiner,
760 Stuttgart.

761 Mejdahl, V., 1979. Thermoluminescence dating: Beta-dose attenuation in quartz grains.
762 Archaeometry 21, 61-72.

763 Messager, E., Lordkipanidze, D., Kvavadze, E., Ferring, C.R., Voinchet, P., 2009. Palaeoenvironmental
764 reconstruction of Dmanisi site (Georgia) based on palaeobotanical data. Quaternary International
765 223-224, 20-27.

766 Messager, E., Lebreton, V., Marquer, L., Russo-Ermolli, E., Orain, R., Renault-Miskovsky, J.,
767 Lordkipanidze, D., Despriée, J., Peretto, C., Arzarello, M., 2011. Palaeoenvironments of early
768 hominins in temperate and Mediterranean Eurasia: new palaeobotanical data from Palaeolithic key-
769 sites and synchronous natural sequences. Quaternary Science Reviews 30, 1439-1447.

770 Meyers, P.A., Teranes, J.L., 2001. Sediment organic matter In: Last, W.M., Smol, J.P., (eds.), Tracking
771 Environmental Change Using Lake Sediments 2. Kluver Academic Publishers, Dordrecht, pp. 239-269.

772 Montpéreux, F. Dubois de, 1842. Reise um den Kaucasus, zu den Tscherkessen, Abchasen, nach
773 Kolchis, Georgien, Armenien und in die Krim, Teil 1, Darmstadt.

774 Morton, R., Gelfenbaum, G., Jaffe, B., 2007: Physical criteria for distinguishing sandy tsunami and
775 storm deposits using modern examples. Sedimentary Geology 200, 184–207.

776 Murray, A.S., Wintle, A.G., 2000. Luminescence dating of quartz using an improved single aliquot
777 regenerative-dose protocol. Radiation Measurements 32, 57-73.

778 Murray, A.S., Wintle, A.G., 2003. The single aliquot regenerative dose protocol: potential for
779 improvements in reliability, Radiation Measurements 37, 377-381.

780 Narimanidze, E., Brückner, H., 1999. Survey on metal contamination of agricultural soils in Georgia,
781 Land Degradation and Development 10, 467-488.

782 Nikolaishvili, D.A., Elizarbarashvili, N., Meladze, G.G., 2011. Evaluation of degree of landscape's
783 anthropogenic transformation (landscapes of Georgia). Procedia Social and Behavioral Sciences 19,
784 547-555.

785 Nikolaishvili, D., Salukvadze, E., Bolashvili, N., Chikhradze, N.G., Matchavariani, L., 2015.
786 Anthropogenic transformation of landscapes of Colchis lowland and surrounding foothills. Earth
787 Sciences 4, 68-72.

788 Nikolaeva, E., Walter, T.R., Shirzaei, M., Zschau, J., 2014. Landslide observation and volume
789 estimation in central Georgia based on L-band InSAR. Natural Hazards and Earth System Sciences 14,
790 675-688.

791 North, C.P.; Davidson, S.K., 2012. Unconfined alluvial flow processes: recognition and interpretation
792 of their deposits, and the significance for palaeogeographic reconstruction. Earth Science Reviews
793 111, 199- 223.

794 Panin, N., Popescu, I., 2007. The northwestern Black Bea: climatic and sea-level changes in the Upper
795 Quaternary. In: Yanko-Hombach V., Gilbert, A.S., Panin, N., Dolukhanov, P.M. (eds.), The Black Sea
796 Flood Question: Changes in Coastline, Climate, and Human Settlement. Springer, New York, pp. 387-
797 404.

798 Papuashvili, R., 2002. Settlement below peat layer in Ontskoshia, Dziebani. Journal of the Centre for
799 Archaeological Research of the Georgian Academy of Sciences 10, 35-38. (in Georgian).

800 Pirazzoli, P.A., 1991. World atlas of Holocene sea-level changes. Elsevier Oceanogr. Ser. 58,
801 Amsterdam, London, New York, Tokyo.

802 Pirazolli, P.A., 1996. Sea-level Changes. The Last 20,000 Years. Wiley and Sons, Chichester.

803 Prescott, J.R., Hutton, J.T., 1994. Cosmic ray contributions to dose rates for luminescence and ESR
804 dating: Large depths and long-term time variations. Radiation Measurements 23, 497-500.

805 Preusser, F., Andersen, B., Denton, G.H., Schlüchter, C., 2005. Luminescence chronology of Late
806 Pleistocene glacial deposits in North Westland, New Zealand. Quaternary Science Reviews 24, 2207-
807 2227.

808 Reimer, P.J., Bard, E., Bayliss, A., Beck, J.W., Blackwell, P.G., Bronk Ramsey, C., Buck, C.E., Cheng, H.,
809 Edwards, R.L., Friedrich, M., Grootes, P.M., Guilderson, T.P., Haflidason, H., Hajdas, I., Hatté, C.,
810 Heaton, T.J., Hogg, A.G., Hughen, K.A., Kaiser, K.F., Kromer, B., Manning, S.W., Niu, M., Reimer, R.W.,
811 Richards, D.A., Scott, E.M., Southon, J.R., Turney, C., van der Plicht J., 2013. InterCal13 and Marine 13
812 radiocarbon age calibration curves 0-50,000 years cal BP. Radiocarbon 55, 1869-1887.

813 Ryan, W.B.F., 2007. Status of the Black Sea Flood Hypothesis. In: Yanko-Hombach, V., Gilbert, A.S.,
814 Panin, N., Dolukhanov, P.M. (eds.), The Black Sea Flood Question. Springer, Dordrecht, pp. 63-88.

815 Ryan, W.B.F., Pitman, W.C., Major, C.O., Shimkus, K., Moskalenko, V., Jones, G.A., Dimitrov, P., Görür,
816 N., Sakınç, M., Seyir, H.I., Yüce, H., 1997. An abrupt drowning of the Black Sea shelf. Marine Geology
817 138, 119–126.

818 Ryan, W.B.F., Major, C.O., Lericolais, G., Goldstein, S.L., 2003. Catastrophic Flooding of the Black Sea.
819 Annu. Rev. Earth Plant. Sci. 31, 525-554.

820 Shatilova, I., Rukhadze, L., Mchedlishvili, N., 2010. The stages of development of vegetation climate
821 of Western Georgia during the Middle and Late Pleistocene. Proceedings of the Natural and
822 Prehistoric Section 2, Georgian National Museum, Tbilisi, Georgia, pp. 75-81.

823 Stewart, H.B, 1958. Sedimentary reflections on depositional environments in San Miguel Lagoon,
824 Baja California, Mexico. Bulletin of the American Association of Petroleum Geologists 42, 2567-2618.

825 Stuiver, M., Reimer, P., 1993. Extended ^{14}C data base and revised calib 3.0 ^{14}C age calibration
826 program. Radiocarbon 35. 215-230.

827 Svanidze, Z., Gogichaishvili, B., Turkadze, G., Alpaidze, V., 2008. Dynamics of heavy metal distribution
828 in soils and natural waters of mining and metallurgical regions. Ecocity World Summit 2008
829 Proceedings, San Francisco, CA, USA, pp. 1-12.

830 Suchodoletz, H. von, Menz, M., Kühn, P., Sukhushvili, L., Faust, D., 2015. Fluvial sediments of the
831 Algeti River in southeastern Georgia – An archive of Late Quaternary landscape activity and stability
832 in the Transcaucasian region. Catena 130, 95-107.

833 Sun, D., Bloemenda, J., Rea, D.K., Vandenberghe, J., Jiang, F., An, Z., Su, R., 2002. Grain-size
834 distribution function of polymodal sediments in hydraulic and aeolian environments, and numerical
835 partitioning of the sedimentary components. Sedimentary Geology 152, 263-277.

836 Tucker, M., 1996. Methoden der Sedimentologie. Ferdinand Enke, Stuttgart.

837 Turney, C., Canti, M., Branch, N., Clark, P., 2005. Environmental Archaeology: Theoretical and
838 Practical Approaches. Routledge, London.

839 Vezolli, G., Garzanti, E., Vincent, S., Andò, S., Carter, A., Resentini, A., 2014. Tracking sediment
840 provenance and erosional evolution of the western Greater Caucasus. Earth Surface Processes and
841 Landforms 39, 1101-1114.

842 Vött, A., 2007. Relative sea level changes and regional tectonic evolution of seven coastal areas in
843 NW Greece since the mid-Holocene. Quaternary Science Reviews 26, 894-919.

844 Vött, A., Schriever, A., Handl, M., Brückner, H., 2007. Holocene palaeogeographies of the central
845 Acheloos River delta (NW Greece) in the vicinity of the ancient seaport Oiniadai. Geodinamica Acta
846 20, 241-256.

847 Wallinga, J., Murray A., Wintle A., 2000. The single-aliquot regenerative-dose (SAR) protocol applied
848 to coarse-grain feldspar. Radiation Measurements 32, 529-533.

849 Wintle, A.G., 1973. Anomalous fading of thermo-luminescence in mineral samples. Nature 245, 143-
850 144.

851 Yanko-Hombach, V., Gilbert, A.S., Dolukhanov, P., 2007. Controversy over the great flood hypotheses
852 in the Black Sea in light of geological, paleontological, and archaeological evidence. Quaternary
853 International 167-168, 91-113.

854 Yılmaz, A., Adamia, S., Yılmaz, H., 2013. Comparisons of the suture zones along a geotraverse from
855 the Scythian Platform to the Arabian Platform. Geoscience Frontiers 5 (6), 855-875.

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Figure captions

Fig. 1: Overview of the research area with location of coring sites and trenches. They are arranged in transects A (blue line) and B (yellow line). Prominent features are the areas of agricultural use on the levees of the rivers Rioni and Khobistsqali. They are in great contrast to the open grasslands, peat bogs and swamp forests. Figure based on Esri basemap and www.naturalearthdata.com.

Fig. 2: Facies interpretation, granulometry, geochemistry and ^{14}C age estimates of the sediment core KUL 3. It represents the stratigraphy of the landward part of the coastal barrier complex, consisting of a (sub-) littoral base, semi-terrestrial back-barrier facies, and aeolian cover sands (deposits of the foredune ridges).

Fig. 3: Facies interpretation, granulometry, geochemistry and age-depth model of the sediment core KUL 7 (caption see Fig. 2). It represents the stratigraphy of the central coastal lowlands, and comprises the entire range of facies: possibly shallow marine deposits at the base (see discussion in chapter 5.1), followed by an interdigitating of lagoonal, semi-terrestrial and alluvial strata.

Fig. 4: Facies interpretation, granulometry, geochemistry and age-depth model of the sediment core PAPO 2 (caption see Fig. 2). The stratigraphy, which is typical for the three southernmost cores, shows river sediments covered by lagoonal deposits. According to the age-depth model, the siltation of the lagoon took place ~2000 years later than at the site of KUL 7.

Fig. 5: Transect A crossing the research area from W to E (position in Fig. 1). While KUL 1, 2 and 3 belong to core group 1 representing the evolution of the coastal barrier complex, KUL 12 belongs to core group 3, which represents the evolution in the centre of the swamps. Cores KUL 7 and KUL 11 are part of core group 4 (see text for further explanations).

Fig. 6: Transect B crosses the research area in S-N direction (position in Fig. 1; legend in Fig. 5). The southernmost cores (SHAV 2, PAPO 3, PAPO 2) belong to core group 1; they are characterized by strong fluvial influence in the lower parts. KUL 9 and KUL 6 represent the evolution in the central part of the swamps, dominated by fine-grained lagoonal mud without intercalation of sand or peat. KUL 7, 10 and 5 belong to group 4 which contains several peat layers, but no major fluvial of alluvial input.

Fig. 7A: Principal component analysis (PCA) of the samples taken from the master cores KUL 3, 7 and PAPO 2. Eight PCs were estimated. The spatial distribution is based on components 1 and 2.with a variance of 36.9 and 19.8% (further variances: PC 3: 14.3%, PC 4: 10.5, PC 5: 8.4%, PC 6: 6%, PC 7: 0%). For the colour code of the samples see Fig. 5.

7B: Ternary plot of mean grain size, sorting and Ca/K ratio, plotted for all samples of the three master cores. For the colour code of the samples see Fig. 5.

7C: Mineralogical compositions of the samples KUL7/36 and KUL 7/45. Qu = quartz, Mu = muscovite, Al = albite, Ca = magnesium-calcite. The obvious difference is that KUL 7/45 contains magnesium-calcite.

Fig. 8: Compilation of ^{14}C-dated samples from the Kolkheti lowlands and their relative position to the local sea level. Horizontal bar shows ^{14}C dating result (2 sigma). The arrows pointing up/down indicate that sea level was higher/lower than the dated sample. In case of paralic peat (facies E; KUL 7/44, KUL 3/10, KUL 3/9), the sea-level range was narrowed to ±0.5 m (see text for further explanations). The data set is compared to sea-level curves from the Taman Peninsula in SW Russia, with the sites of SEM = Semebratnee (Brückner et al., 2010), GOL = Golobitskaya (Kelterbaum et al., 2011), and DZHI = Dschiginka (Fouache et al., 2012). The differences are due to different tectonic settings. For the study area, a continued and rather moderate rise in sea level for the last eight millennia is assumed.

904 **supplementary data**

905 *Appendix 1: Sample preparation for luminescence (IRSL) dating*

906 The following sample preparation and luminescence measurements were carried out in the Cologne
907 Luminescence Lab (CLL) under subdued red light. After HCl, H_2O_2 and $Na_4O_7P_2$ treatments to remove
908 carbonates, organics and clay contents, gravity separation was carried out using heavy liquid
909 solutions (sodium polytungstate; ρ_1 = 2.68 g/cm^3, ρ_2 = 2.62 g/cm^3, ρ_3 = 2.58 g/cm^3) in order to
910 separate coarse-grained (100-200 µm) quartz and feldspar. All measurements were carried out on an
911 automated Risø TL/OSL DA 20 reader equipped with a ^{90}Sr/^{90}Y beta source delivering 0.08 Gy/sec.

912 Quartz is commonly used to date Holocene deposits due to its ubiquity and a generally stable signal
913 for dating young deposits. However, in this study, potassium-rich feldspars were preferred for dating
914 because of inappropriate luminescence properties of the quartz (e.g. low signal intensities).
915 Nonetheless, feldspar can exhibit anomalous fading, which is defined as a signal loss due to leakage
916 of electrons at room temperature. This can cause an age underestimation (see Wintle, 1973;
917 Lamothe and Auclair, 1999; Balescu et al., 2001; Huntley and Lamothe, 2001). To include the effect of
918 anomalous fading in the age calculation, several sub-samples were tested and measured to observe a
919 possible decrease in the IRSL response over time, following Auclair et al. (2003). Indeed, a signal loss
920 was detected in all sub-samples. The g-values (1.9 and 1.3 %/decade) were calculated for the
921 samples KUL 1/4 and KUL 2/2. Thus, the fading was corrected by using the procedure proposed by
922 Huntley and Lamothe (2001), as the equivalent dose was derived using linear parts of the growth
923 curve. The resulting ages are given both with uncorrected and fading corrected values. The fading
924 correction for sample KUL 1/2 refers to the g-value of KUL 1/4, and the fading correction for sample
925 KUL 2/1 was conducted using the g-value of KUL 2/2 (Tab. 2).

926 For the feldspar measurements an infrared stimulation (880 ± 80 nm) and an interference filter (410
927 nm) were used. All measurements were carried out following the single aliquot regenerative dose
928 protocol (SAR) after Murray and Wintle (2000, 2003) and Wallinga et al. (2000). Due to the preheat
929 plateau test, which indicated no plateau building at higher temperatures, a preheat temperature of
930 180 °C was chosen. Each aliquot was thus preheated at 180 °C (according to combined preheat-
931 plateau-dose-recovery-tests) for 10 sec and then measured for 300 sec at 50 °C (IRSL 50 protocol
932 after Wallinga, 2000).

933 To calculate the annual dose rate derived from the decay of lithogenic radionuclides in the
934 sediments, the concentrations of uranium, thorium and potassium were determined by laboratory
935 high-resolution γ-spectrometry using the software DRAC by Durcan et al. (2015). The conversion
936 factors of Adamiec and Aitken (1998), and α- and ß- attenuation factors of Bell (1979) and Mejdahl
937 (1979) were applied. Following the approach of Prescott and Hutton (1994) the cosmic dose rate was
938 calculated. An internal radiation of 12.5 ± 0.5 % potassium (Huntley and Baril, 1997) and an α-
939 efficiency factor of 0.07 ± 0.02 (Preusser et al., 2005) were assumed for feldspars.

940