

2009-04

Record of Cenozoic sedimentation from the Amanos Mountains, Southern Turkey: Implications for the inception and evolution of the ArabiaEurasia continental collision

Boulton, SJ

<http://hdl.handle.net/10026.1/3044>

10.1016/j.sedgeo.2009.01.008

Sedimentary Geology

Elsevier BV

All content in PEARL is protected by copyright law. Author manuscripts are made available in accordance with publisher policies. Please cite only the published version using the details provided on the item record or document. In the absence of an open licence (e.g. Creative Commons), permissions for further reuse of content should be sought from the publisher or author.

1 **NOTICE: this is the author's version of a work that was accepted for publication in**
2 **Sedimentary Geology. Changes resulting from the publishing process, such as peer review,**
3 **editing, corrections, structural formatting, and other quality control mechanisms may not be**
4 **reflected in this document. Changes may have been made to this work since it was submitted**
5 **for publication. A definitive version was subsequently published in Sedimentary Geology, 216,**
6 **29-47 (2009). doi:10.1016/j.sedgeo.2009.01.008**
7

8
9 **Record of Cenozoic sedimentation from the Amanos Mountains, Southern Turkey:**
10 **implications for the inception and evolution of the Arabia-Eurasia continental collision**
11

12 Sarah J. Boulton

13
14 School of Earth, Ocean and Environmental Sciences, University of Plymouth, Drakes Circus,
15 Plymouth, PL4 8AA, UK.

16 E-mail - sarah.boulton@plymouth.ac.uk; fax: 01752 233117
17
18

19 **Abstract**

20 The sedimentary succession of the southern Amanos Mountains, bordering the eastern margin of the
21 Karasu Rift in south central Turkey, provides a record of environmental change from the Eocene
22 (Lutetian) to the Upper Miocene (Tortonian) that charts the final evolution of the northern margin of
23 the Arabian plate prior to and during continental collision. Eocene shallow-marine carbonates
24 (Hacıdağı Formation) are interpreted as the youngest unit of the Arabian passive margin succession
25 deposited on a northwards facing carbonate ramp. Subsequent deformation and uplift took place
26 during the Oligocene represented by folding of the Eocene and older strata. This is interpreted to be
27 the result of initial continental collision between Arabia and Eurasia. Unconformably overlying the
28 Eocene limestone are Lower Miocene conglomerates, sandstones and palaeosols up to 150 m thick
29 (K1c1 Formation). These were deposited in a range of marginal marine settings consisting of alluvial

30 fan/fan delta facies, flood plain as well as basinal facies. Subsequently, during the Middle Miocene,
31 local patch reefs developed in restricted areas (Kepez Formation) followed by Upper Miocene
32 sediments (Gökdere Formation) composed of relatively deep water hemipelagic marl, with clastic
33 interbeds, which represent a transgression during this period. The Upper Miocene becomes sandier
34 upwards, this records the regression from the relatively deep water facies to coastal sediments. Water
35 depth gradually became shallower until during Pliocene time the area became continental in nature.
36 By the Quaternary rifting had resulted in the development of the Karasu Rift with active alluvial fans
37 along the margins and braided rivers depositing coarse conglomerates in the axial zone. These
38 conglomerates are interbedded with basaltic lava flows that resulted from the region extension across
39 the area. This research shows that initial continental collision occurred in this area after the Lutetian
40 (40.4 Ma) and before the Aquitanian (23.03 Ma) supporting the hypothesis that the southern
41 Neotethys Ocean closed during the Late Eocene to Oligocene. This was a time of climatic change
42 including the onset of southern hemisphere glaciation, in which the closure of the southern Neotethys
43 may have had played an important role.

44

45 Key words: Neogene, carbonate ramp, alluvial fan, continental collision, Dead Sea Fault, Neotethys,
46 Eocene-Oligocene boundary.

47 **Cite as:** Boulton, S. J., 2009. Record of Cenozoic sedimentation from the Amanos Mountains,
48 Southern Turkey: implications for the inception and evolution of the Arabia-Eurasia continental
49 collision. *Sedimentary Geology*. 216, 29-47. doi:10.1016/j.sedgeo.2009.01.008

50

51 **1 Introduction**

52 It is now generally accepted that the geology of southern Turkey records evidence for the
53 evolution and closure of the Southern Neotethys Ocean and the timing of the collision between

54 Arabia and Anatolia (Sengor and Yilmaz, 1981; Robertson and Dixon, 1984; Yilmaz et al., 1993;
55 Robertson, 2000; Robertson et al., 2004; Robertson et al., 2006). However, there is still much
56 debate on when the northward subduction of Arabia beneath Anatolia ceased and when the closure
57 of the southern Neotethys and subsequent continental collision actually took place (Hall, 1976;
58 Aktaş and Robertson, 1984; Yilmaz et al., 1993; Beyarslan and Bingöl; 2000). There are three main
59 alternative theories, with collision occurring either during: 1) the Late Cretaceous (Karig and Kozlu,
60 1990; Kozlu 1997; Beyarslan and Bingöl, 2000); 2) in the Late Eocene (Vincent et al., 2007; Allen
61 and Armstrong, 2008) or 3) during the Oligocene to Early Miocene (Aktaş and Robertson, 1984;
62 Yilmaz et al., 1993; Robertson, 2000; Robertson et al., 2004; Robertson et al., 2006) along a suture
63 that runs through SE Turkey (Bitlis Suture; Fig. 1) and into Iran (Zagros Suture; Fig. 1).

64 The Tertiary evolution of the northern margin of the southern Neotethys has attracted much
65 attention (e.g. Hall, 1976; Aktaş and Robertson, 1984; Yilmaz et al., 1993; Beyarslan and Bingöl,
66 2000; Robertson, 2000; Robertson et al., 2004; Robertson et al., 2006), but there has been less focus
67 on the coeval evolution of the Arabian margin in the west. Research has been carried out on
68 Cenozoic sediments found within the Bitlis suture zone. These sediments are thought to have been
69 deposited within basins assumed to be representative either of a peripheral foreland basin (Şengör
70 and Yilmaz, 1981; Kelling, 1987) or of small transtensional basins (Karig and Kozlu, 1990; Kozlu,
71 1997). Additionally, there is some work focussing on the evolution and role of Cenozoic basins on
72 the Arabian platform that are today > 100 km from the suture zone (Hatay Graben, Turkey; Boulton
73 et al., 2006; Boulton and Robertson, 2007: Nahr El-Kabir half-graben, Syria; Hardenberg and
74 Robertson, 2007). Conversely, much of the work on the Arabian margin has concentrated on the
75 Zagros region in Iran and Iraq, 1000s of kilometres to the east (Hessami et al., 2001; Agard et al.,
76 2005; Vincent et al., 2005) and therefore, may not be representative of regions to the west if
77 continental collision was diachronous.

78 In this paper, I will present the first modern descriptions and interpretation of the Eocene to
79 Late Miocene sedimentary sequence from the southern end of the Amanos Mountains, ~ 50 km

80 south of the suture zone (Figs. 1, 2). This study focuses on the area around the towns of Kırıkhan,
81 Belen and Serinyol in the Hatay Province of southern Turkey where Cenozoic strata outcrop in the
82 Amanos Mountains due to uplift on the flanks of the Plio-Quaternary Karasu Rift. New
83 sedimentological and petrological data are presented for these rocks based upon a recent
84 redefinition of the stratigraphic framework (Boulton et al., 2007). This allows new
85 palaeoenvironmental interpretations to be made and implications drawn for the palaeogeographic
86 evolution of the northern Arabian plate margin during the final stages of continental collision and
87 the subsequent development of a peripheral foreland basin on the Arabian plate.

88

89 **2 Geological Framework**

90 The DSFZ system forms the boundary between Arabia and Africa (Fig. 1), accommodating the
91 difference in motion between the two plates through sinistral strike-slip motion. The DSFZ trends ~
92 N-S from the Red Sea, in the south, to the junction with the East Anatolian Fault Zone (EAFZ) near
93 Kahramanmaraş in southern Turkey, in the north. The DSFZ developed, in the south, during the
94 Middle Miocene (dated as <20 Ma, Lyberis, 1988; 18 Ma, Garfunkel and Ben Avraham, 1996) with
95 the slip rate calculated as ~7 mmyr⁻¹ (Garfunkel et al., 1981).

96 The Karasu Rift forms the northernmost segment of the Dead Sea Fault Zone (DSFZ), located
97 primarily in the Hatay Province of Turkey, trending northwards from the Amik Plain (Fig. 1). To
98 the south of the Amik Plain, the Gharb Rift forms the southwards continuation of the DSFZ; while
99 to the east another structure, the Hatay Graben, trends NE-SW to the present Mediterranean coast.
100 The Karasu Rift can be subdivided into three segments (Rojay et al., 2001); northern, central and
101 southern. This paper will focus on the western margin of the southern segment of the Karasu Rift
102 from Kırıkhan in the north to Serinyol in the south.

103 The Karasu Rift is bounded by two main faults the Amanos Fault Zone (AFZ) in the west
104 and the East Hatay Fault (EHF) in the east (Fig. 1). Slip rate estimates of the AFZ from offset

105 lavas range from 0.3 mmyr^{-1} (Arger et al., 2000, based upon data from Çapan et al., 1987), 2 – 15
106 mmyr^{-1} (Rojay et al., 2001), 1 – 1.6 mmyr^{-1} (Yurtmen et al., 2002) to 4 mmyr^{-1} (Tatar et al., 2004).
107 There are few estimates for the slip rates of the EHF, amongst which those of Tatar et al., (2004)
108 who derived a figure of 4 mmyr^{-1} for the EHF. Although motion on these faults is dominantly
109 sinistral strike-slip there is a normal component of motion that has lead to the development of the
110 Karasu Rift and, which caused the uplift of the Amanos Mountains to the west of the rift. The
111 Amanos Mountains are composed of a core of metamorphosed Precambrian rocks, Palaeozoic
112 sediments, ophiolite and Cenozoic sedimentary cover, whereas the rift fill is poorly exposed recent
113 alluvium with interbedded lavas (Fig. 2).

114 Pioneering geological research in the study area was undertaken by Dubertret (1939, 1953),
115 followed by the first subdivision of the Palaeozoic rocks of the Amanos Mountains by Dean and
116 Krummenacher (1961) and of the complete stratigraphic sequence by Atan (1969). Geological
117 mapping of the Amanos Mountains as a whole was undertaken by Schwan (1971), Ishmawi (1972),
118 Janetzko (1972) and Lahner (1972) but these works generally lacked widespread correlation and
119 palaeontological age constraints. The 1980s saw an increased interest in the area by the Turkish
120 Petroleum Corporation (detailed list in Dean et al., 1986). Although the majority of this work is still
121 unpublished, the work of Guney (1984) presents important micropalaeontological biozonation of
122 Cenozoic formations. Piskin et al., (1986) published a new geological map for the Hatay region,
123 with a brief synopsis of the sedimentary units based mostly on the earlier work of Atan (1969). In
124 the same year, Dean et al., (1986) presented a revised stratigraphic scheme for the Palaeozoic
125 sediments of the southern Amanos area (around Kırıkhan), which did not include the Cenozoic
126 stratigraphy. Further geological mapping in the Kırıkhan and Belen areas was undertaken by Kop
127 (1996) and Dokumaci (1997).

128 It is evident that the sedimentary cover has only received rudimentary attention, in
129 comparison to the detailed analysis and interpretation of the Upper Cretaceous ophiolite (e.g.,
130 Çogulu, 1974; Delaloye et al., 1980; Piskin et al., 1986), the structural geology of the Karasu Rift

131 (e.g., Saroglu et al., 1992; Rojay et al., 2001; Westaway and Arger, 2001; Adiyaman and
132 Chorowicz, 2002; Over et al., 2002; Yurtmen et al., 2002; Westaway, 2003; Tatar et al., 2004;
133 Akyuz et al., 2006) and the nature of the Karasu basalts (Çapan et al., 1987; Parlak et al., 1998;
134 Rojay et al., 2001; Yurtmen et al., 2002; Tatar et al., 2004).

135

136 **3 Cenozoic sediments of the Karasu Rift**

137 The Cenozoic sediments of the Southern Karasu Rift (Table 1) occur predominantly around the
138 towns of Kırıkhan, on the western edge of the Karasu Rift, and Belen in the Amanos Mountains
139 (Fig. 2). The strata range from Eocene to Late Miocene in age, the Eocene strata are folded and
140 faulted (Boulton and Robertson, 2008) and unconformably overlain by relatively undeformed
141 Neogene sedimentary rocks. This sedimentary cover overlies the Hatay/Kızıldağ Ophiolite that was
142 emplaced southwards during the Maastrichtian (Robertson, 2002). A smaller outcrop of Cenozoic-
143 aged sediment is preserved on the edge of the Amik Plain around the town of Serinyol, to the north
144 of Antakya and southwest of Kırıkhan. This area lies in the southernmost part of the Karasu Rift
145 within a re-entrant of the ophiolite. A sequence of Palaeocene-Eocene to Upper Miocene
146 sedimentary rocks is exposed in this small area (Fig. 3). In this section sedimentary descriptions and
147 interpretations of the Cenozoic units are given, which are then used to build a model of
148 palaeogeographic change in the study area allowing comparisons to other areas and an evaluation of
149 major tectonic events.

150

151 **3.1 Eocene limestone - Hacıdağı Formation.**

152 There are large exposures of Eocene limestone along the Belen and Kırıkhan roads to Antakya. The
153 limestone (Hacıdağı Formation) varies from fine-grained wackestone and packstone to rudstone.
154 Bedding is thin (0.1-0.2 m thick) with sharp bases and tops and often normally graded, bases of
155 some beds are erosional and many are laterally discontinuous. Additionally, bedding is disrupted in

156 places where slumping has occurred or was incipient (Figs. 4a, b). Data from fold hinges and
157 bedding in slumped horizons indicates a general $\sim 000^\circ - 020^\circ$ for the down-slope direction (Fig. 5).
158 Planar lamination is common, and there is abundant bioturbation, where burrows can be seen to cut
159 and disrupt the laminations in some beds. There are numerous large benthic foraminifera, often
160 concentrated in lags at the bases of the beds. Planktic foraminifera are also reported from these
161 limestones such as *Globorotalia velascoensis* (Cushman), but benthic species are the most diverse
162 (*Nummulites* sp., *Discocyclina* sp., *Orbitolites* sp., *Alveolina* sp., *Assilina* sp. (Atan, 1969)) dating
163 the formation to the Lutetian (Boulton et al., 2007). Chert is common, forming dark grey to black,
164 elongated nodules parallel to the bedding planes. Occasionally there are horizons of angular- to
165 well-rounded clasts, composed of limestone and serpentinite. Overlying these thinly bedded
166 limestones is a laterally extensive bed, ~ 10 m thick, that has an erosive base. This is a matrix-
167 supported conglomerate with sub-rounded clasts of fine-grained, white limestone (< 70 mm) in a
168 matrix of non-fossiliferous white nummulitic wackestone/floatstone.

169 Along the Belen-Antakya road section, the basal contact between the Hacıdağı Formation
170 and the underlying ophiolite (sheeted dyke complex) is generally sharp and does not appear to have
171 any basal conglomerate or other lithological changes associated with it, and is therefore interpreted
172 as a faulted contact. The upper boundary of the formation can be observed on the main road just
173 before Belen. The top of the Eocene limestone is eroded and bedding is sub-vertical. Overlying
174 basal Miocene sediments contain well-rounded clasts of Eocene limestone indicating an erosional
175 contact.

176 Eocene limestones are also exposed along a riverbed near Serinyol (Fig. 6b). Here the
177 formation (Fig. 3) directly overlies the eroded upper surface of the ophiolite, along a disconformity.
178 The base of the sequence is composed of hard white limestone (wackestone-packstone) containing
179 large benthic foraminifera (*Nummulites*, *Discocyclina*) and common chert nodules. Within this
180 basal limestone there is a large, laterally discontinuous, conglomerate horizon (Fig. 4c), the base of
181 which is irregular and cuts down to the top of the ophiolite in the north. This conglomerate is

182 poorly sorted and clast supported, composed entirely of sub-angular to sub-rounded, limestone and
183 chert clasts (Table 1) up to 0.9 m in size. The matrix is sandy and contains *Amphistegina* sp.,
184 *Operculina* sp., *Rotalia viennotti*, *Quinqueloculina* sp., *Globigerina* sp., *Miogypsina* sp., *Textularia*
185 sp., *Lepidocyclina* sp., and unidentifiable foraminifera of the Rotaliina. There are also fragments of
186 *Lithophyllum*.

187 Above the conglomerate, there is a return to limestone (fine-grained sparite). Bedding is thin
188 and irregular; the upper surfaces of the beds occasionally exhibit current ripples. Chert nodules are
189 very common and *Nummulites* and other large benthic foraminifera are occasionally observed in
190 dense accumulations. Within this sequence there are several clast supported conglomerate horizons;
191 the clasts are angular to sub-rounded, <0.5 m in size and composed of limestone and chert. There is
192 also a thick (2.5 m) rudstone bed that has an irregular base that cuts down into the underlying beds.
193 *Nummulites* and *Discocyclina* exhibit a rough parallel alignment. Additionally, there is a small
194 change in bedding orientation above and below this bed.

195 A conglomerate horizon is observed near the abandoned village of Kanlidere (Fig. 6a).
196 Thin-bedded, wackestone with chert nodules is exposed in the valley bottom. The top of the
197 formation is a 7-10 m thick, clast-supported conglomerate, clasts are sub-angular to sub-rounded
198 composed of limestone and chert (Table 3), with a maximum clast size of 0.3 m. Although
199 generally poorly sorted there are horizons within the conglomerate that exhibit better sorting.

200 3.1.1 Interpretation

201 The characteristic planar beds, which fine-upwards and exhibit parallel lamination, suggest
202 that these limestones were deposited from low density turbidity currents consisting of the T_{abe}
203 divisions of the classical Bouma sequence (Bouma, 1962). The T_{cd} divisions of the Bouma
204 sequence are generally absent, suggesting that partial flow separation may have taken place
205 resulting in incomplete sequences.

206 Some horizons are distorted and broken; these were interpreted as incipient slumps as the
207 bedding was broken but significant offset in the bedding planes had not occurred (Fig 4a). Other

208 horizons show more obvious evidence of slumping including a spectacular isoclinal fold (Fig. 4b).
209 The orientation of these structures indicates the presence of a northward-dipping slope (Fig. 5). The
210 thin-bedded facies are commonly capped by a thick bed of laterally extensive, matrix-supported
211 conglomerate. The clast size, shape and overall texture suggest the conglomerate was probably lain
212 down by a powerful a debris flow (Nilsen, 1982).

213 Poorly sorted, clast-supported conglomerates were observed at Serinyol and Kanlidere (Fig.
214 6). The clast-supported nature suggests deposition was not by a debris-flow process but that it was
215 deposited from a hyperconcentrated sediment flow. In addition, several conglomerate horizons are
216 present near Serinyol. The main conglomerate bed cuts out the basal part of the underlying bedded
217 Eocene limestone and sits directly on top of the underlying ophiolite. It is probable that this
218 conglomerate represents a channel fill; however, it was not possible to determine the channel
219 orientation. Foraminifera identified in the matrix indicate derivation from a shelf environment.
220 Near the top of the sequence there is a widespread but slight angular (1-2°) discordance within the
221 formation, directly below this there is a thick bed of Nummulitic rudstone with an erosive base; this
222 may be the result of scouring, possibly following a tectonic event. Only one conglomerate horizon
223 was observed at Kanlidere. The lenticular nature of these clast-supported conglomerates suggests
224 that these could be channel-fill deposits or scour fills.

225 The abundance of large benthic foraminifera and low abundance of planktic forms indicates
226 relatively shallow water depths. The presence of *Nummulites* sp., in particular indicates water
227 depths of 20 – 75 m (Saller et al., 1993) although reworking to deeper water is likely as the
228 foraminifera are often found in concentrated lags.

229 Therefore, the biotal evidence combined with evidence for sediment instability and
230 movement (slumps, turbidites and channel fills) suggests that during the Eocene in this northern
231 area, carbonates were being deposited on a slope, orientated approximately northwards. The slope
232 was unstable generating debris flows and turbidity currents and was cut into by large channels that
233 were in-filled by coarse-grained sediments.

234

235 **3.2 Lower Miocene sandstone and conglomerates - Kıcı Formation.**

236 The base of the Kıcı Formation rests unconformably on the Eocene at the type section near
237 Kurtisoguksu (K; Fig. 2); as is the case throughout the area as Oligocene sediments are absent. A
238 thick-bedded, dominantly matrix-supported conglomerate unit, 50-60 m thick, is present at the base
239 of the section (Fig. 7). Clasts are up to 1 m in size, angular to sub-rounded and composed
240 dominantly of carbonate, although there are some basaltic and red sandstone (probably from the
241 underlying Palaeozoic strata) clasts present. Above this basal conglomerate there is a sequence of
242 coarse-grained purplish-red sandstone. Beds are 0.3-3 m thick with sharp bedding plane contacts.
243 Grain-size is generally very coarse but also conglomeratic and mudstone horizons are present.
244 Sandstone beds often exhibit normal grading, contain “floating” rounded pebbles (< 20 mm in
245 general) and pebble stringers that contain predominately serpentinite clasts (derived from the
246 underlying ophiolite). In general, serpentinite is the main constituent of the sandstones in the Kıcı
247 Formation (Table 3). Sedimentary structures are common, such as parallel lamination, planar cross-
248 bedding (< 0.1 m high foresets) and bioturbation (Fig. 8). It should be noted however, that reliable
249 palaeocurrent measurements could not be determined due to the poor nature of the exposure.

250 Conglomerate beds are clast-supported, with well-rounded clasts of a generally ophiolitic
251 composition; rare limestone clasts are also present. The conglomerates are either found at the base
252 of sandstone beds or form laterally discontinuous lenses. There is also an interval with 1.75 m of
253 mudstone exposed. The bottom 1 m is composed of thin interbeds of white and pale lilac-coloured
254 chalk and black mudstone, which is overlain by dark grey mudstone containing roots and plant
255 material.

256 The Lower Miocene succession is also observed to overlie the Eocene limestones above an
257 erosional unconformity near Gökdere (Fig. 4d); however, no basal conglomerate is present at this
258 location, only 1.5 m of mixed breccia and red mudstone (Fig. 9). Above this there is ~5 m of
259 coarse-grained boundstone, composed mostly of large fragments of algal material (Rhodophyta;

260 *Lithophyllum* sp., *Corallina* sp.: Dasycladaceae; *Halimedia* sp.), other bioclastic (Foraminifera;
 261 *Textularia* sp., *Spiroloculina* sp., *Globigerina* sp., *Amphistegina* sp., *Quinqueloculina* sp.,
 262 *Triloculina* sp., *Globorotalia* sp., including *G. menardii*, *Rotalia viennotti* and *Miogypsina* sp) and
 263 some siliciclastic material. Upwards, the sequence is dominated by coarse litharenite and
 264 conglomerate. Conglomerates are matrix-supported and composed of rounded clasts of dominantly
 265 ophiolite material (Table 2). Limestone clasts are also present, which occasionally show evidence of
 266 boring. Clasts are well-rounded and the size varies between beds. Rounded “floating” pebbles are
 267 very common within the sandstone beds.

268 Higher in the section, sedimentary structures become more common and parallel lamination,
 269 trough and planar cross-bedding, pebble imbrication and bioturbation are present. Although, current
 270 indicators are present most are unsuitable for measurement, three measurements on imbricate
 271 pebbles were measured (080°, 090°, 120°) suggesting an easterly flow. Additionally, there are
 272 occasional mudstone beds, generally black in colour with a shaley fabric, with abundant plant
 273 material and rootlets in some horizons. The top of the sequence is poorly exposed but it appears
 274 that there is an upward transition to the overlying Kepez Formation. The coarse litharenite is light
 275 purple to red due to the high content of altered serpentinite. Thin-sections reveal that basaltic clasts
 276 and polycrystalline quartz are also common.

277 To the southwest, at Kanlidere, the Lower Miocene is markedly different, there ~40 m of
 278 Lower Miocene sediments outcrop. Fine- to medium-grained, poorly sorted, massive brown
 279 sandstone is exposed (Fig. 6a), containing unidentified gastropods, fragmented and reworked
 280 *Corallina* sp and *Lithophyllum* sp., algae, and some foraminifera (*Triloculina* sp., *Neoalveolina* sp.,
 281 *Spiroloculina* sp., and unidentified Rotaliina). Interbedded with the massive sandstone there are
 282 three conglomerate horizons (Fig. 6a). These conglomerates are clast-supported, poorly sorted with
 283 well-rounded to sub-rounded clasts up to 0.4 m in size (Fig. 10a). The matrix is composed of fine-
 284 grained, brown sandstone, similar to the rest of the exposure and the clasts have a mixed
 285 composition of limestone and ophiolite (Table 2). Many of the limestone clasts exhibit *Lithophaga-*

286 like borings. The top of the formation is conformable with the overlying limestone (Middle
287 Miocene Kepez Formation).

288 The Lower Miocene, near Serinyol, is different again. The sediments here are composed of
289 30 - 40 m of red and brown mottled mudstones with abundant caliche nodules and palaeosol
290 horizons (Fig. 6b). However, individual beds cannot be distinguished due to the heavily weathered
291 and diagenetically altered nature of the sediments. No conglomerate was observed in this section.

292 3.2.1 *Interpretation*

293 Although there is no dating evidence for this formation, the stratigraphic position below the
294 Kepez Formation suggests an Early Miocene age and foraminifera identified in the basal bioclastic
295 facies observed at Gökdere, confirm a Miocene age for the base of the formation. The sedimentary
296 characteristics of the K1c1 Formation can be broadly differentiated into three facies associations;
297 alluvial fan, braided stream and shoreline.

298 The presence of both matrix- and clast-supported conglomerates at the base of the type
299 section, is suggestive of both debris-flow and sheet flood processes. Using standard nomenclature
300 (Miall, 1978; 1985; 1996) these sediments can be classified as Gmm and Gcm facies, indicating a
301 SG facies association, interpreted as coarse alluvial fan sediments. The disorganised fabric, large
302 clast sizes and matrix-supported intervals indicate deposition on a debris flow dominated alluvial
303 fan (Postma, 1986; Blair and McPherson, 1994). There is no fossil material present, consistent with
304 a non-marine origin.

305 In contrast, the basal sediments consisting of 1.5 m of breccia and palaeosols at Gökdere,
306 probably formed through exposure and weathering of the underlying limestone. This is overlain by
307 microbial boundstone; fossil content indicates a shallow-marine origin for this material with some
308 reworking and abrasion causing fragmentation of the bioclasts. The encrusting nature of the
309 Rhodophyta (red algae) caused biological binding of the carbonate and suggests a high-energy
310 shallow-marine setting (Wright and Burchette, 1996), such as the edge of an algal reef.

311 The upper part of the Lower Miocene is composed of conglomerate and coarse litharenites
312 with occasional mudstone horizons. However, the type section is a fining-upwards sequence (Fig.
313 7), whereas a coarsening-up sequence is identified at Gökdere (Fig. 9). Conglomerate horizons
314 generally occur at the base of sandstone beds or as laterally discontinuous lenses at the type section.
315 By contrast, there is a higher proportion of conglomerate present at Gökdere. The limestone clasts
316 locally show evidence of *Lithophaga*-like boring, indicating that some pebbles were reworked in a
317 shallow marine setting before being incorporated into the these conglomerates, as *Lithophaga* sp.
318 live in the littoral zone of marine coasts.

319 The cross-bedded sandstones with basal lags of conglomerate seen at the type location are
320 suggestive of lateral and downstream accretion macroforms and possibly sandy bedforms,
321 suggesting possible deposition from a braided coarse-grained to sandy bedload river (Miall, 1996).
322 The fine-grained sediments consisting of soft dark grey to black mudstone indicates high organic
323 matter content indicative of water-logged conditions. The presence of rootlets and plant material
324 suggests non-marine conditions and colonisation by plants, indicating deposition on a flood plain,
325 abandoned channel, or marsh adjacent to the active channel.

326 By contrast, the sandstone exposed at Kanlidere contains bioclastic material including
327 coralline algae, gastropod and bivalve fragments, and both benthic and planktic foraminifera,
328 indicating a marine (possibly a shallow-marine peri-reefal) environment. The massive nature of the
329 sandstone suggests that it has undergone intense bioturbation. Three clast-supported conglomerate
330 beds are interbedded with the sandstone (Fig. 6a), suggesting a stream-flow or tractional reworking
331 and winnowing and deposition in a coastal setting. Some of the limestone clasts have borings, also
332 indicating reworking in a shallow-marine environment prior to deposition. The features described in
333 these sediments may indicate deposition in the lower shoreface. The mudstones near Serinyol,
334 interpreted as a palaeosol succession (Fig. 6b), indicate this area was emergent and undergoing soil
335 formation during the Early Miocene.

336 In addition to the shoreface deposits at Kanlidere, there is evidence of marine influence (i.e.
337 basal boundstone) at Gökdere, suggesting that this location was proximal to the coast unlike the
338 type section and represents a regressive sequence. This could imply that these locations represent a
339 lateral transition of alluvial fan-braided/meandering river-deltaic environments. Therefore, these
340 facies associations indicate a range of depositional environments along a coarse-grained gravel rich
341 coast as evidenced by the interaction of marine and alluvial processes. This coarse-grained coast
342 appears to evolve over time from an alluvial fan delta to a braid delta (Orton, 1988) reflecting the
343 change in the feeder system from a point sourced alluvial fan to a braided river system (feeder
344 system types A and B of Postma, 1990). It is not clear whether this fan-delta system is a low
345 gradient shelf-type, shallow water delta or a slope-type, deep water delta (Postma, 1990) as the pro-
346 delta sediments have not been identified.

347 The composition of the sandstone of the K1c1 Formation is dominated by serpentine, basalt
348 and radiolarian chert clasts and there is very little matrix or cement present. This indicates that the
349 sediment source was probably the underlying ophiolite and related rocks. In contrast, the
350 conglomerate horizons at Kanlidere were dominantly derived from the sedimentary cover (mainly
351 limestone and chert). Fine-grained sediments contain abundant quartz and muscovite; these
352 minerals could be extrabasinal as mica, especially, is uncommon in the local basement rocks.

353

354 **3.3 Middle Miocene - Kepez Formation: limestone.**

355 The Kepez Formation is poorly exposed only at a few locations. Near Kepez Hill (adjacent to
356 Gökdere village), the Middle Miocene is a small exposure of rubblely limestone. This is composed
357 of shallow-marine debris, such as fragments of oncolite, coral, bivalves, gastropods and echinoids.
358 Additionally, a large amount of coralline (mainly poritid corals) debris is strewn about the hillside
359 in this area.

360 Near Kırıkhan, there is a small outcrop of fine-grained crystalline wackestone containing
361 fragmented coralline algae, foraminifera and echinoids. This bed is variable in thickness (2-3 m)

362 and has an uneven basal surface. This overlies an extremely poorly exposed soft marly limestone.
363 By contrast, 10-15 m of Middle Miocene limestone is well exposed at Kanlidere (Fig. 6a). The
364 basal beds are marly wackestone and pass upwards into hard packstone. There is abundant
365 bioclastic debris, bivalve fragments, bryozoa, echinoids, small gastropods, coral and oncolites.

366 The Middle Miocene exposure near Serinyol is irregular and of a variable thickness (2 to 6
367 m) composed of recemented rubblely material. Blocks are ~10cm in size, angular and clast-
368 supported. The limestone is rich in bioclastic material, such as *Pecten*, *Ostrea*, poritid corals and
369 gastropods (Fig. 10b). Underlying this material there is an irregular basal bed of limestone rich in
370 fragmented bivalves.

371

372 3.3.1 Interpretation

373 These limestones are very poorly exposed and the bioclastic material is fragmented;
374 therefore, there is little information on which to base an environmental reconstruction. Generally,
375 the bioclastic packstone indicates formation in shallow-marine conditions, which possibly
376 accumulated slightly offshore as the fragmentation of the bioclastic material indicates that it was
377 reworked. Also, the large blocks of limestone seen in the north are characteristic of reef talus.

378 The large coral fragments and the rubblely nature of the limestone, near Serinyol, indicate
379 that this may also be reef talus, confirming the presence of shallow-marine reefs. As this facies is
380 observed in a number of different locations (Fig. 3) and as large fringing reefs are uncommon in the
381 Mediterranean during the Miocene (Franseen et al., 1996), it is most likely that this material was
382 derived from small patch reefs.

383

384 3.4 Upper Miocene mudstone with sandstone interbeds - Gökdere Formation.

385 The type section of the Upper Miocene at Gökdere is composed of fine-grained, grey marl
386 (Figs. 10c, 12; Table 4). Near the base of the formation are thin (50 mm) interbeds of fine-grained
387 micaceous litharenite (Table 3). Some of these thin beds exhibit parallel laminations and contain

388 plant fragments, small gastropods and ostracods, but marl sampled for faunal studies were barren.
389 Upwards the proportion of sand increases; sandstone beds become thicker (< 0.5 m) and more
390 abundant (Fig. 10d). Sandstone beds are usually found in packets with a significant thickness of
391 marl between. The sandstone is laterally discontinuous with sharp upper and lower bedding
392 surfaces. There are an abundance of sedimentary structures present in the sandstones, including
393 small channel structures, parallel lamination, cross-bedding, ripples, rip-up clasts, load casts, and
394 current-aligned plant material.

395 The proportion of sand continues to increase upwards until bedding thickness exceeds 1 m;
396 these beds continue to be interbedded with marl. Near the top of the exposed sequence thin micritic
397 carbonate horizons are interbedded with the sandstone and marl (Fig. 11). The sandstone weathers
398 orange and sedimentary structures are uncommon; although mud rip-up clasts and bioturbation were
399 observed. There are many elongate carbonate-cemented concretions. In thin section these massive
400 litharenites are composed of a range of lithic fragments (dominantly serpentinite and limestone
401 clasts), quartz, muscovite, biotite and rare glauconite with some clay matrix and sparry calcite
402 cement. Additionally, there are fragments of transported algal clasts as well as (transported?)
403 planktic (*Globigerina* sp., *Globigerinoides* sp.) and benthic foraminifera (*Amphistegina* sp., and
404 indeterminate Rotaliina). Occasional oyster beds are present, where individual *Ostrea* specimens
405 can exceed 0.2 m in length. *Turitella* gastropods and plant material are present in adjacent beds.
406 Thin carbonate beds exhibit parallel lamination and, in thin-section, layers of disarticulated
407 ostracods valves apparently of only two taxa. The ostracod species are undetermined but *Cyprideis*
408 *seminulum* (Reuss) and *Cyprideis anatolica* (Bassiovini) have been reported from this formation by
409 Kop (1996).

410 Around the town of Belen and to the south, the Upper Miocene is well exposed, here
411 composed of interbedded sandstone and marl/mudstone. The mudstone is very fine-grained,
412 variable in colour and forms the majority of the succession. The litharenite/calcarenite is fine- to
413 coarse-grained; beds are normally graded and micaceous. Bed thickness is generally < 0.5 m, but

414 most beds are < 0.1 m thick; laterally, these beds are discontinuous, tending to form discrete packets
415 within the marl. Sedimentary structures are common; parallel laminations, ripples, cross-
416 laminations were all observed. Fallen blocks reveal that the bases of the beds have various sole
417 marks (formed by erosion and bioturbation). Fossils are not generally present but *Ostrea* fragments
418 were identified and plant material is quite common.

419 The basal Upper Miocene sediments near Serinyol are grey marl, containing small bivalves
420 (undetermined) and foraminifera (*Globigerinoides trilobus* (Reuss); *Orbulina suturalis*
421 (Brönniman); *Orbulina bilobata* (d'Orbigny); Boulton et al., 2007) and also fragments of larger
422 bivalves and polyzoans. Upwards, the colour changes to brownish. There are occasional horizons
423 with parallel laminations but there are no major lithological variations. The upper part of the
424 formation, includes packages of sandstone beds, these are 10-20 m thick and formed from thin,
425 irregular beds of medium-grained calcarenite, separated by a similar thickness of marl. Ripples,
426 planar cross-lamination, tepee structures and rip-up clasts are present. Some fossil material is
427 present, mostly as fragmented bivalves.

428 Palaeocurrent measurements from this formation were based generally on ripples, sole
429 marks and, at the type section, from flow-oriented plant debris. Although there is some spread in
430 palaeocurrent orientation, in general they indicate a southerly to westerly flow (Fig. 12).

431

432 3.4.1 Interpretation

433 The Upper Miocene forms a coarsening-upward sequence. This sequence is interpreted to
434 represent a progradational shoreline succession, characterised by a shallowing and coarsening-
435 upwards sequence from marine offshore muds to silt and sand facies of the shoreface (Reading &
436 Collinson, 1996). The basal sediments dominated by marl deposition are interpreted as basinal
437 mudstones.

438 Higher in the succession the proportion of sand increases, probably relating to shallowing
439 and regression. The laterally discontinuous, generally massive beds with erosional bases are likely

440 to be channel fills. Whereas, the more laterally continuous beds that show a wider range of
441 sedimentary structures, such as parallel laminations, cross-lamination, ripples and rip-up clasts, are
442 interpreted as storm generated current deposits due to the similarities to the current-modified
443 turbidite model of Myrow and Southard (1996).

444 At the type section (Fig. 11), the sequence continues to coarsen upwards and biodiversity is
445 low in some beds with only one species of *Turitella* and occasional large *Ostrea* present. This low
446 diversity fossil assemblage could suggest a stressed, possibly brackish water, environment but may
447 also be the result of the preferential dissolution of aragonite post-deposition. However, the presence
448 of *Ostracoda* probably *Cyprideis* sp., in thin carbonate horizons is indicative of brackish water,
449 lagoonal or lacustrine environments of very shallow water, < 10 m in depth (Neale, 1988).
450 Conversely, a more diverse assemblage, albeit reworked, is found in the massive sandstones
451 indicating that this material was transported from fully marine conditions. These massive beds
452 could have been deposited in the lower shoreface where bioturbation may have obliterated
453 sedimentary structures or again they may be tempestite deposits. Palaeocurrent analysis indicates a
454 general direction of sediment transport to the west/southwest (Fig. 12), suggesting shore-oblique
455 currents consistent with storm generated deposition (Myrow and Southard, 1996).

456 At locations other than the type section, these upper coarse sandstones and thin carbonate
457 horizons are absent. This may be because erosion has removed the upper part of the Miocene in
458 these other localities or that the shoreline was further into the basin and in the subsurface at the
459 present time.

460 The marls are composed dominantly of argillaceous lime mud but also contain significant
461 amounts of quartz (Table 4). Quartz is not expected to be present in high volumes if erosion of the
462 underlying ophiolite is the main sediment source. This suggests that some detritus was being
463 sourced from other rocks types, possibly from quartzite exposed in the core of the Amanos
464 Mountains north of Kırıkhan (e.g., Sadan, Sosink, Seydişehir Formations), or from even more
465 distant sources, such as the Taurides. Muscovite is present in some samples from the Kıcı and

466 Gökdere Formations and is likely to have an extrabasinal source, possibly from metamorphic rocks,
467 as micaceous rocks are not common in the immediate area.

468 **4 Discussion**

469 **4.1 Evolution of the north-western Arabian margin**

470

471 The Hatay (Kızıldağ) and Baer-Bassit Ophiolites were emplaced during the Maastrichtian
472 (Al-Riyami et al., 2002), southwards onto the Arabian margin. Rising eustatic sea-level (Miller et
473 al., 2005) possibly combined with isostatic regional subsidence following ophiolite emplacement
474 resulted in a widespread marine transgression across the Arabian platform throughout Palaeocene
475 and Early Eocene times. Directly to the south of the study area, in the Nahr El-Kabir region of
476 Syria, Early – Middle Eocene limestones (Fig. 13) are characteristic of open-shelf conditions with
477 evidence of westward shallowing during the Middle Miocene (Hardenberg and Robertson, 2007).
478 Similarly, to the southeast of the study area, in the present Hatay Graben, Lutetian sedimentation
479 was characterised by a marine transgression from intertidal to shallow open marine conditions
480 (Boulton and Robertson, 2007). However, in the Belen- Kırıkhan area facies are dominated by
481 turbidites and northward flowing slumps, indicating a depositional position on the outer ramp or
482 ramp slope of the Neotethys margin (Fig. 14a). This field evidence is in agreement with current
483 palaeogeographic models of the region for that time (e.g., Meulenkamp and Sissingh, 2003).

484 Chert is common in these facies and similar facies have been observed in Eocene
485 carbonates in Hatay, Turkey (Boulton and Robertson, 2007), Syria (Hardenberg and Robertson,
486 2007), Israel (Rosenfeld and Hirsch, 2005) and onshore and offshore Cyprus (Robertson, 1998).
487 The presence of diagenetic chert is generally attributed to high diatom and radiolarian productivity
488 either due to upwelling on the Neotethys margin (Boulton and Robertson, 2007) or due to a marine
489 connection between the Mediterranean and the Indian Ocean (Hardenberg and Robertson, 2007).

490 Following the deposition of the Eocene carbonates there was a period during which folding
491 and uplift took place (Boulton and Robertson, 2008). It is likely that this deformation explains the
492 Oligocene hiatus, which formed during the initial continental collision between Eurasia and Arabia.
493 Evidence collected here indicates that the collision took place sometime between the Lutetian
494 (youngest platform carbonates in the area) and the Aquitanian (assumed age of the oldest sediments
495 above the regional unconformity surface). Such deformation has also been recorded for the Hatay
496 basin to the southeast (Boulton and Robertson, 2007) and recognised to north on the conjugate
497 margin seen in the east-central Taurides (Karig and Kozlu, 1990), but was not mentioned by
498 Hardenberg and Robertson (2007) as having effected coeval strata in northern Syria. Interestingly,
499 although strata of Oligocene-age are not present in the study area or in the Hatay Graben, strata
500 from this epoch are present elsewhere to the south, in Syria and Jordan, indicating that the hiatus in
501 sedimentation during the Oligocene was confined to the northernmost margin of the Arabian plate.
502 This may indicate that the collisional front was located near to the present-day Bear-Bassit area of
503 northern Syria. It has also been suggested (Hempton, 1985; Sharland et al., 2004) that non-
504 deposition in these northern areas of the Arabian plate may have resulted from local uplift during
505 reactivation of structural lineaments related to the Syrian Arc, but it may also be significant that the
506 Oligocene was a period of low eustatic sea-level (Miller et al., 2005).

507 During the Early-Middle Miocene conglomerates and sandstones were deposited (Fig. 14b).
508 Basal sediments were deposited in an alluvial fan setting. Alluvial fans typically form in regions of
509 active deformation where a hinterland with steep relief is separated from a lower relief basin by a
510 rapid change in slope gradient (Heward, 1978). This may indicate the presence of an active fault at
511 this time, but there is no supporting evidence for syn-sedimentary tectonic movement so it is
512 suggested that the fan formed as a result of erosion of mountains rapidly uplifted by initial
513 continental collision. Upwards there is a transition to a braid delta system, which prograded east-
514 south eastwards, into a marine embayment in the Amik Plain area (Fig. 14b). Clast sizes are
515 significantly smaller higher in the section, this decrease through time could be the result of lowered

516 relief. Palaeosols exposed near Serinyol are interpreted as the lateral, terrestrial equivalent to these
517 sediments deposited in a floodplain environment.

518 The dominant clast lithology of serpentinite in the K1c1 Formation indicates that the
519 sediment source was the ophiolite emplaced in the Maastrichtian. However, the presence of
520 limestone clasts identified as being similar to the Eocene facies, indicates that the young carbonates
521 had already been uplifted and were being eroded; this probably took place during the Oligocene.
522 The K1c1 Formation can be correlated with Early Miocene braided-river sediments identified in the
523 Hatay basin to the southeast (Balyatağı Formation; Boulton and Robertson, 2007) that flowed
524 northwards from the Baer-Bassit Massif, Syria (Fig. 13). This contrasts with a marine signature for
525 some parts of the Early Miocene sediments in the Kırıkhan area, although a brief marine incursion
526 in the Early Miocene has been recorded from the Hatay basin as well (Boulton and Robertson,
527 2007). This implies that regionally there was a marine incursion in the earliest Miocene with a
528 regressive trend occurring through the Early and Middle Miocene leading to continental deposition
529 in the Hatay basin and coastal environments to the north.

530 By contrast, Early Miocene sediments to the north of Kırıkhan, in the suture zone of the
531 orogeny, are significantly different. These are mainly deep water turbidites that thicken southwards
532 (Lice Formation: Aktas and Robertson, 1984; Karig and Kozlu, 1990; Robertson et al., 2004; Gül,
533 2006). Sole marks in those Aquitanian sediments indicate an easterly flow direction along an east-
534 west orientated basin in the northeast, whereas in the Iskenderun basin the dominant flow direction
535 was to the southwest (Karig and Kozlu, 1990).

536 To the south at this time the Nahr El-Kabir half graben developed separating the Baer-Bassit
537 Massif in the northeast from the main Arabian shelf to the south and east (Hardenberg and
538 Robertson, 2007). Sedimentation during this time was dominated by marine carbonates (Fig. 13).
539 This indicates that Baer-Bassit, Hatay Basin, Amanos range and parts of the southern Karasu Rift
540 represented a topographic high during the Early Miocene between the shallow marine carbonate
541 platform to the south and the foreland (Lice and Iskenderun Basins) to the encroaching collisional

542 front to the north. This is in agreement with Boulton and Robertson (2007) who proposed that the
543 Early Miocene sediments in the Hatay Basin represented erosion of the flexural forebulge created
544 by tectonic loading of the subducting slab accentuated by reactivation of basement structures.

545 The Middle Miocene saw the development of localised patch reefs in the study area (Fig.
546 14c), which pass laterally and vertically up into a coarsening-up sequence of marl to sandstone. The
547 presence of patch-reefs indicates suitable conditions for coral growth; however, these conditions
548 were short lived possibly due to the influx of fine-sediment stifling coral growth. The overlying
549 marl sequence has been dated as Serravallian to Messinian in age (Kozlu, 1982) and the patch reefs
550 as Langhian–Serravallian (Kozlu, 1982). The marls are marine in the south near Serinyol, however,
551 the presence of *Ostrea* sp., *Cyprideis* sp., and *Turitella* sp., suggests a restricted marine to brackish
552 water-environment in the north, at the top of the succession, indicating a shallowing in water depth
553 to the northeast (Fig. 14d).

554 This sequence shares many similarities to the sediments in the Hatay area, although faulting
555 related to the formation of the Hatay Graben appears to have initiated during the Middle Miocene
556 (Boulton et al., 2006) the area was still part of the wider foreland basin to the thrust front (Boulton
557 and Robertson, 2007). Sedimentation during the Langhian and Serravallian was dominated by
558 shallow-marine peritidal carbonate deepening up into outer ramp carbonates and then marls over
559 time possibly due to flexural subsidence relating to foreland loading (Boulton and Robertson,
560 2007). Patch reefs were identified in the northwest of that area (near Kesecik, Boulton and
561 Robertson, 2007; their figure 1), close to the area discussed in this paper indicating that the study
562 area was also part of the shallow underfilled foreland basin of the collisional zone. It is possible that
563 subsidence was greater in the southeast due to the effect of local normal faulting superimposed on
564 regional subsidence; whereas the absence of normal faulting in areas to the north reduced overall
565 subsidence rates and the resulting accommodation space.

566 However, the marls of the Hatay and Karasu sequences differ in composition with higher
567 concentration of quartz and mica present in the sediments around Kırıkhan, whereas mica is

568 uncommon in the Hatay Graben and quartz concentrations are much lower (Boulton, 2006). This
569 implies that the Hatay Graben was more distal from these detrital sediment sources than the
570 Kırıkhan area. The quartz could have been eroded locally but there are no micaceous basement
571 rocks in the vicinity suggesting that this material was transported from a greater distance. The
572 nearest outcropping micaceous rocks are to the north of the suture zone, in the Berit region
573 (described in detail by Robertson et al., 2006) some 150 km from the study area, there mica schists
574 and granites outcrop. This suggests that by the Late Miocene, sediment eroded from north of the
575 suture was being transported across the suture zone into the foreland basin that developed in front of
576 the leading edge of the collision.

577 By contrast, to the south, in present day Syria, the main development of the Nahr El-Kabir
578 half-graben took place in the Middle Miocene (Hardenberg and Robertson, 2007). There is a Late
579 Serravallian thin chalk horizon present, but there are no sediments of Tortonian age present (Fig.
580 13). Hardenberg and Robertson (2007) explain this as due to local tectonics influencing
581 sedimentation. Unlike the Hatay and Nahr El-Kabir areas, evidence for syn-sedimentary faulting
582 has not been observed in Middle-Late Miocene sediments around Kırıkhan, Belen or Serinyol.

583 Messinian evaporites are not present in this area, although they have been identified to the
584 north in the Iskenderun Bay (Boulton, 2006), in the Hatay Graben (Boulton et al., 2006; Boulton
585 and Robertson 2007) and to the south in Syria (Hardenberg and Robertson, 2007). It is not clear
586 whether evaporites were either not deposited in the area, were deposited and subsequently eroded,
587 or are buried in the subsurface.

588 Pliocene sediments have not been identified around Kırıkhan, Belen or Serinyol, existing
589 micropalaeontological dating indicates the youngest sediments near Serinyol are of Late Miocene
590 age (Boulton et al., 2007). Pliocene sediments are present to the southeast in the Hatay Graben
591 (Boulton et al., 2006) and to the south in the Nahr El-Kabir Graben (Hardenberg and Robertson,
592 2007) but both of these were tectonically active basins during the Neogene, whereas the Kırıkhan
593 area was not. It is likely that regional Pliocene-Recent uplift, related to continued convergence

594 between Arabia and Anatolia, induced terrestrial deposition earlier in the Kırıkhan-Karasu region,
595 compared to areas to the south and west (Fig. 14e).

596 In addition to regional uplift, the DSFZ propagated northwards during the Pliocene.
597 Transtension resulted in strike-slip and extensional components of deformation (Boulton, in
598 review), the extensional component of deformation caused flank uplift and basin floor subsidence
599 leading to the formation of the present topographic graben. The lack of dated Pliocene sediments
600 and a Quaternary age for the rift fill suggests that significant topography did not develop until the
601 Late Pliocene. Subsequently, up to ~350 m of Pleistocene river gravel (mostly unexposed) have
602 accumulated within the Karasu Rift (Rojay et al., 2001). Moreover, some 11 inactive alluvial fans
603 are interbedded with basalts that have yielded dates ranging from 1.57 – 0.05 Ma (Rojay et al.,
604 2001; Yurtmen et al., 2002; Tatar et al., 2004). These basalts are associated with volcanic necks and
605 have been attributed to block rotations and extension of the Karasu Rift (Tatar et al., 2004).

606

607 **4.2 Timing of continental collision and implications**

608 The Amanos Mountains are located at the westernmost interface between an extensive mobile
609 belt to the east that has been deformed and uplifted following the closure of the southern Neotethys
610 Ocean and consequent Arabia-Eurasia collision, and an extensive area to the west that has yet to
611 undergo full continental collision (Mediterranean Basin). The study area is also located to the south
612 of the Bitlis suture zone (Fig. 1). Constraining the timing of deformation, initial uplift and
613 subsequent evolution of the southern Karasu Rift allows this area to be integrated into the broader
614 geotectonic framework and enhances our understanding of the orogenic evolution in this sector of
615 the Alpine-Himalayan chain.

616 In the East, south of the Zagros suture zone, collision has been shown to have commenced in
617 the Late Eocene to Oligocene. Agard et al., (2005) constrained the timing of collision to between
618 35 and 25-23 Ma; after the last intrusion of mafic igneous material related to arc volcanism and
619 prior to the onset of Late Oligocene/Early Miocene sedimentation. This is corroborated by the work

620 of Hassami et al., (2001) on progressive unconformities in the Zagros that indicate deformation
621 began in the Late Eocene. Further to the west in northern Iraq, terrestrial clastics dated to the Late
622 Eocene have been inferred to represent sub-aerial uplift and erosion of the northeastern edge of the
623 Arabian plate by this time (Dhannoun et al., 1988). Collectively, these data corroborate the
624 findings from the Karasu Rift and Hatay Graben, where folded Eocene and older strata underlie an
625 extensive hiatus of Oligocene indicating that compressional deformation along the north Arabian
626 margin was taking place from the Late Eocene onwards.

627 Similarly, evidence from the north of the suture zone in the Caucasus and Caspian Basin
628 indicates that the onset of collision also took place during the Late Eocene – Oligocene (Patton,
629 1993; Vincent et al., 2005; 2007) due to the presence of deformed and eroded Eocene strata
630 unconformably overlain by clastics of presumed Oligocene age, olistostromes and compressional
631 deformation observed in the subsurface. Late Eocene uplift has also been recorded in northern Iran
632 (Alborz), where a Middle Eocene basin was inverted by the Early Oligocene (Alvavi, 1996; Guest
633 et al., 2006).

634 In southern Turkey, north of the Bitlis suture zone, structural, sedimentological and
635 stratigraphical studies have determined that the initial collision took place between the Late Eocene
636 (Yilmaz, 1993) and Early Miocene time. Robertson et al., (2004; 2006) proposed that diachronous
637 oblique subduction continued throughout the Eocene and that olistostromes indicate that the latest
638 stages of subduction and initial collision took place in the Oligocene to Early Miocene. In many
639 central Anatolian basins, sedimentation continued through the Late Eocene but was terminated by a
640 basal Oligocene unconformity that is present in nearly all basins (e.g., Şarkışla basin, Gökten, 1986;
641 Sivas basin, Dilek et al., 1999; Ulukisla basin, Clark and Robertson, 2005; Tuzgölü basin, Görür et
642 al., 1989). In addition, latest Eocene folding is also reported for these same areas indicating that
643 deformation propagated rapidly northwards into the interior of the Anatolian plate, as well as
644 propagating rapidly northwards into the interior of the Eurasian Plate (Vincent et al., 2005; 2007).
645 Alternatively, deformation initiation took place over a wide area.

646 The evidence present here, supports that from the region and shows that widespread
647 deformation took place during the Late Eocene to Oligocene from Iran to western Anatolia. This
648 indicates that the closure of the southern Neotethys and initial continental collision took place
649 almost simultaneously along the entire frontal sector of the Arabian Plate with associated
650 deformation propagating rapidly into the hinterland. This suggests that rather than the collision
651 being diachronous in nature, it was broadly synchronous along the leading edge of the Arabian
652 Plate.

653 Interestingly, the Eocene to earliest Oligocene was also a period of rapid expansion of the
654 Antarctic continental ice sheet (Zachos et al., 2001); the closure of the Tethys Ocean in conjunction
655 with other oceanographic changes, such as the widening of the North Atlantic Ocean (Zachos et al.,
656 2001), would seem to be a significant factor in the climatic changes that occurred at the Eocene-
657 Oligocene boundary.

658 **5 Summary and Conclusions**

659 Deposition of the Cenozoic sediments exposed in the Amanos Mountains, was dominantly
660 controlled by subsidence related to continental collision taking place to the north and concomitant
661 foreland basin formation south of the suture zone. The study area represents a Late Cretaceous to
662 Eocene north-facing continental shelf at the southern Neotethys passive continental margin.. The
663 Late Eocene to Oligocene is absent in the area due to erosion or non-deposition of sediments
664 indicating that during some, if not all, of this period the area was uplifted and eroded. This uplift is
665 attributed to the flexure of the crust and southward migration of the flexural bulge resulting from
666 loading of the Arabian lithosphere due to continental closure of the Neotethys to the north. Recent
667 research (e.g., Patton, 1993 Vincent et al., 2005; 2007; Allen and Armstrong, 2008) indicates that
668 initial continental collision appears to have taken place nearly simultaneously along the leading
669 edge of Arabia. Although, high resolution studies need to be undertaken to confirm this as the Late

670 Eocene to end Oligocene is a period of some 20 Ma and diachronism of the continental collision
671 cannot be excluded completely.

672 Continental sedimentation during the Early Miocene reflects erosion of uplifted areas due to
673 regional deformation resulting from the final closure of the Neotethys and suture tightening along
674 the Misis-Andırın lineament. The flexural bulge passed to the south of the area and was shedding
675 material northwards into the foreland basin; however, the proto-Amanos Mountains appear to
676 already have developed into a topographic high and were additionally shedding sediment
677 southwards. Marine transgression during the Middle to Late Miocene resulted in the deposition of
678 localised patch reefs and clastic sediments were deposited into local depocentres (i.e., Hatay basin,
679 Amik basin) with palaeocurrents directed to the south and west, indicating that the palaeoslope was
680 orientated towards the developing Hatay Graben and not northwards towards the thrust front.

681 Regional uplift combined with a general regressive trend resulted in continental conditions by
682 the latest Miocene/Pliocene that continue to the present day. Transtension, related to the
683 northwards propagation of the DSFZ, resulted in the formation of the Karasu Rift during the
684 Pliocene. The extensional component of deformation created accommodation space and fluvial
685 conglomerates accumulated within the axial zone of the rift. These sediments are interbedded with
686 lavas that resulted from localised extension and block rotations in the rift floor.

687

688 **Acknowledgements**

689 My thanks go to Malcolm Hart for foraminiferal identifications and to Alastair Robertson and
690 Matthew Watkinson who commented on earlier versions of this manuscript. Additionally, the
691 suggestions of the editor Gert Jan Weltje, Johan ten Veen and an anonymous reviewer greatly
692 improved this manuscript. This work was initiated during a NERC PhD studentship at the
693 University of Edinburgh and completed at the University of Plymouth with funding from the Royal
694 Society.

695

696 **References**

697 Adiyaman, O., Chorowicz, J., 2002. Late Cenozoic tectonics and volcanism in the north western
698 corner of the Arabian Plate: a consequence of the strike-slip Dead Sea Fault zone and the lateral
699 escape of Anatolia. *Journal of Volcanology and Geothermal Research*, 117, 327-345.

700

701 Agard, P., Omrani, J., Jolivet, L., Mouthereau, F., 2005. Convergence history across Zagros (Iran):
702 constraints from collisional and earlier deformation. *International Journal of Earth Sciences*, 94,
703 401-419.

704

705 Aktas, G., Robertson, A.H.F., 1984. The Maden Complex, S E Turkey: evolution of a Neotethyan
706 continental margin. In: Dixon, J.E., Robertson, A.H.F. (Eds.). *The Geological Evolution of the*
707 *Eastern Mediterranean*. Geological Society, London, Special Publications, 17, 375-402.

708

709 Akyuz, H. S., Altunel, E., Karabacak, V., Yalciner C. C., 2006. Historical earthquake activity of
710 the northern part of the Dead Sea Fault Zone, southern Turkey. *Tectonophysics*, 426 (3-4), 281-
711 293.

712

713 Allen, M. B., Armstrong, H. A., 2008. Arabia-Eurasian collision and the forcing of mid Cenozoic
714 global cooling. *Palaeogeography, Palaeoclimatology Palaeoecology*, doi:
715 10.1016/j.palaeo.2008.04.021

716

717 Alavai, M., 1996. Tectonostratigraphic synthesis and structural style of the Alborz mountain
718 system in northern Iran. *Journal of Geodynamics*, 21, 1-33.

719

720 Al-Riyami, K., Robertson, A.H.F., Dixon, J.E, Xenophontos, C., 2002. Origin and emplacement of
721 the Late Cretaceous Baer-Bassit ophiolite and its metamorphic sole in NW Syria, *Lithos*, 65, 225-
722 260.

723

724 Arger, J., Mitchell, J., Westaway, R., 2000. Neogene and Quaternary volcanism of south-eastern
725 Turkey. In: Bozkurt, E., Winchester, J. A., Piper, J. D. A. (eds). *Tectonics and Magmatism of*
726 *Turkey and the Surrounding Area*. Geological Society, London, Special Publications, 173, 459-487.

727

728 Atan, O. R. 1969. Egribucak Kirikhan bölgesindeki ofiyolilerin jeolojisi ve petrografisi [Geology
729 and Petrography of the Kirikhan District Ophiolite]. M.T.A yayini no 150, pp71, Ankara.

730

731 Beyarslan, M., Bingöl, A. F., 2000. Petrology of a super-subduction zone ophiolite (Elazığ
732 Turkey). *Canadian Journal of Earth Sciences*, 37, 1411-1424.

733

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

737

738 Boulton, S. J., Transpression or transtension? Structure of the northern Dead Sea Fault Zone. In
739 review for *Journal of Structural Geology*.

740

741 Boulton, S. J., 2006. Tectonic-sedimentary evolution of the Cenozoic Hatay Graben, South Central
742 Turkey. University of Edinburgh, unpublished PhD thesis, 414pp.

743

- 744 Boulton, S. J., Robertson, A. H. F., 2007. The Miocene of the Hatay area, S Turkey: Transition
745 from the Arabian passive margin to an underfilled foreland basin related to closure of the Southern
746 Neotethys Ocean. *Sedimentary Geology*, 198, 93-124.
- 747
- 748 Boulton, S. J., Robertson, A. H. F., Unlügenç, Ü. C., 2006. Tectonic and sedimentary evolution of
749 the Cenozoic Hatay Graben, Southern Turkey: A two-phase, foreland basin then transtensional
750 basin model. In: Robertson, A.H.F., Mountrakis, D. (eds.), *Tectonic Evolution of the Eastern
751 Mediterranean*. Geological Society, London, Special Publications, 260, 613-634.
- 752
- 753 Boulton, S. J., Robertson, A. H. F., Ellam, R. M., Safak, Ü., Ünlügenç., U. C., 2007 Strontium
754 isotopic and micropalaeontological dating used to redefine the stratigraphy of the Neotectonic
755 Hatay Graben, southern Turkey. *Turkish Journal of Earth Sciences*, 16, 141-179.
- 756
- 757 Bouma, A. H., 1962. *Sedimentology of some Flysch deposits: A graphic approach to facies
758 interpretation*, 168p, Elsevier, Amsterdam.
- 759
- 760 Bull, W. B., 1972. Recognition of alluvial fan deposits in the stratigraphic record. In: Rigby, J.K. &
761 Hamblin, W. K. (eds) *Recognition of ancient sedimentary environments*. Society of Economic
762 Palaeontology and Mineralogy Special Publication, 16, 63-83.
- 763
- 764 Çapan, U. Z., Vidal, P., Cantagrel, J. M., 1987. K –Ar, Sr and Pb isotopic study of Quaternary
765 volcanism in the Karasu valley (Hatay), N-end of the Dead-Sea rift zone in SE-Turkey. *Hacettepe
766 University Earth Sciences*, 14, 165-178.
- 767

- 768 Clark, M., Robertson, A., 2005. Uppermost Cretaceous – Lower Tertiary Ulukisla Basin, south
 769 central Turkey: sedimentary evolution of part of a unified basin complex within an evolving
 770 Neotethyan suture zone. *Sedimentary Geology*, 173, 15 -51.
 771
- 772 Çogulu, E., 1974. Ultrabasic tectonites and layered peridotites of the Hatay area (Turkey). *Bulletin*
 773 *of the Maden Tektik Arama*, 83, 139-147.
 774
- 775 Dean, W. T., Krummenacher, R., 1961. Cambrian trilobites from the Amanos Mountains, Turkey.
 776 *Palaeontology*, 4, 71-81.
 777
- 778 Dean, W. T., Monod, O., Günay, Y., 1986. Lower Palaeozoic stratigraphy in the southern and
 779 central Amanos Mountains, south central Turkey. *Geological Magazine*, 123 (3), 215-226.
 780
- 781 Delaloye, M., Piskin, Ö., Selçuk, H., Vuagnat, M. & Wagner, J-J. 1980. Geological section through
 782 the Hatay ophiolite along the Mediterranean coast, southern Turkey. *Ofiliti*, 52 (3), 205-216.
 783
- 784 Dhannoun, H. Y., Aldabbagh, S. M. A., Hasso, A. A., 1988. The geochemistry of the Gercus Red-
 785 bed formation of Northeast Iraq. *Chemical Geology*, 69, 87-93.
 786
- 787 Dilek, Y., Thy, P., Hacker, B., Grundvig, S., 1999. Structure and petrology of Tauride ophiolites
 788 and mafic dike intrusions (Turkey): implications for the Neotethyan ocean. *Geological Society of*
 789 *America Bulletin*, 111, 1192– 1216.
 790
- 791 Dokumaci, Y., 1997. Belen tektono-stratigrafik incelemesi (Tectono-stratigraphic investigations of
 792 the Belen area). MSc. Thesis, University of Çukurova.
 793

- 794 Dubertret, L. 1939. Sur la genèse et l'âge des roches vertes syriennes. Comptes Rendus de
 795 l'Académie des Sciences, Paris, 209 P763.
 796
- 797 Dubertret, L. 1953. Géologie des roches vertes du nord-ouest de la Syrie et du Hatay (Turquie).
 798 Notes et Mémoires sur le Moyen-Orient, Museum Nationale et histoire naturelle, 5, 5-179, Paris.
 799
- 800 Franseen, E. K., Esteban, M., Ward, W. C., Rouchy, J-M., 1996. Models for carbonate stratigraphy
 801 from Miocene reef complexes of Mediterranean regions. Society for Sedimentary Geology –
 802 Concepts in Sedimentology and Paleontology Volume 5, 391pp
 803
- 804 Garfunkel, Z., Ben-Avraham, Z., 1996. The structure of the Dead Sea Basin. Tectonophysics, 255,
 805 155-176.
 806
- 807 Garfunkel, Z., Zak, I., Freund, R., 1981. Active Faulting in the Dead Sea Rift. Tectonophysics, 80,
 808 1-26.
 809
- 810 Gökten, E., 1986. Palaeocene carbonate turbidites of the Şarkışla region, Turkey—their significance
 811 in an orogenic basin. Sedimentary Geology 49, 143– 165.
 812
- 813 Görür, N., Tüysüz, O., Şengör, A.M.C., 1998. Tectonic evolution of the Central Anatolian Basins.
 814 International Geological Review, 40, 831–850.
 815
- 816 Guest, B., Stockli, D. F., Grove, M., Axen, G. J., Lam, P. S., Hassanzadeh, J., 2006. Thermal
 817 histories from the central Alborz Mountains, northern Iran: Implications for the spatial and temporal
 818 distribution of deformation in northern Iran. Geological Society of America Bulletin, 118, 1507 –
 819 1521.

820

821 Gül, M., 2006. Evolution of the turbidite system in the Kahramanmaraş Basin. Cükürova
822 University, Unpublished PhD thesis.

823

824 Günay, Y. 1984. Amanos daglarının jeolojisi ve Karasu-Hatay grabeninin petrol olanakları. TPAS
825 Hakkari-Saryaj Projesi, TPAO-1984, Ankara.

826

827 Hall, R., 1976. Ophiolite emplacement and evolution of the Taurus suture zone. South-east
828 Turkey. Geological Society of America, Bulletin, 87, 1078-1088.

829

830 Hardenberg, M., Robertson, A.H.F., 2007. Sedimentology of the NW margin of the Arabian plate
831 and the SW-NE-trending Nahr El-Kabir half-graben in northern Syria during latest Cretaceous-
832 Cenozoic. Sedimentary Geology. Doi: 10.1016/j.sedgeo.2007.02.009

833

834 Hempton, M. R., 1987. Constraints on Arabian Plate motion and extensional history of the Red Sea.
835 Tectonics 6, 687-705.

836

837 Hessami, K., Koyi, H. A., Talbot, C. J., Tabasi, H., Shabanian, E., 2001. Progressive
838 unconformities within an evolving foreland fold-thrust belt, Zagros Mountains. Journal of the
839 Geological Society, London, 158, 969 – 981.

840

841 Heward, A.P., 1978. Alluvial fan sequence and mega sequence models: with examples from
842 Westphalian D-Stephanian B coalfields, northern Spain. In: Maill, A.D. (ed) Fluvial
843 Sedimentology. Canadian Society of Petroleum Geologists, Memoirs, 5, 669-702.

844

845 Holmes, A. 1965. Principles of Physical Geology, 2nd Edition. Nelson, London. Pp 1288.

846

847 Ishmawi, R., 1972. Geologie des nordlichen Mittelteils des Amanos-Gebirges zwischen Islahiye
848 und Bahçe (S. Türkei). Geotektonischen Forschungen, 42, 34-63.

849

850 Janetzko, P., 1972. Geologische Untersuchungen an der Ostflanke des sudlichen Amanos-Gebirges
851 zwischen Islahiye und Hassa (Sudtürkei). Geotektonischen Forschungen, 42, 3-33.

852

853 Karig, D. E., Kozlu, H., 1990. Late Palaeogene – Neogene evolution of the triple junction region
854 near Maraş, south central Turkey. Journal of the Geological Society, London, 147, 1023-1034.

855

856 Kelling, G., Gökçen, S. L., Floyd, P. A., Gökçen, N., 1987. Neogene tectonics and plate
857 convergence in the eastern Mediterranean: New data from southern Turkey. Geology, 15, 425-429.

858

859 Kop, A., 1996. Kirikhan ve Kuzeyinin tektono-stratigrafik incelemesi (Tectono-stratigraphic
860 investigations of Kirikhan and its northern area). MSc Thesis, University of Çukurova, Turkey.

861

862 Kozlu, H. 1982. İskenderun Baseni Anadolu nun Kambrien teşekkülleri ve bunların Doğu İran
863 Kambrieni ile mukayesesi. MTA periodical no 66, Ankara.

864

865 Kozlu, H., 1997. Tectono-stratigraphic units of the Neogene basins (Iskenderun, Misis-Andirin) and
866 their tectonic evolution in the eastern Mediterranean region. [PhD Thesis] Natural Science Institute,
867 Cukurova University, Turkey.

868

869 Lahner, L., 1972. Geologische Untersuchungen an der Ostflanke des des mittleren Amanos (SE
870 Türkei). Geotektonischen Forschungen, 42, 64-96.

871

872 Lyberis, N., 1988. Tectonic evolution of the Gulf of Suez and the Gulf of Aqaba. *Tectonophysics*,
873 153 (1-4), 209-220.

874

875 Miall, A. D., 1978. *Fluvial Sedimentology*. Canadian Society of Petroleum Geologists Memoir, 5,
876 859pp.

877

878 Miall, A. D., 1985. Architectural element analysis: a new method of facies analysis applied to
879 fluvial deposits. *Earth-Science Reviews*, 22, 261-308.

880

881 Miall, A. D., 1996. *The Geology of Fluvial Deposits: Sedimentary facies, basin analysis and*
882 *petroleum geology*. Springer-Verlag, Berlin, 582 pp.

883

884 Miller, K. G., Kominz, M. A., Browning, J. V., Wright, J. D., Mountain, G. S., Katz, M. E.,
885 Sugarman, P. J., Cramer, B. S., Christie-Blick, N., Pekar, S. F., 2005. The Phanerozoic record of
886 global sea-level change. *Science*, 310 (5752), 1293-1298.

887 Meulenkamp, J. E., Sissingh, W., 2003. Tertiary palaeogeography and tectonostratigraphic
888 evolution of the Northern and Southern Peri-Tethys platforms and the intermediate domains of the
889 African-Eurasian convergent plate boundary zone. *Palaeogeography, Palaeoclimatology,*
890 *Palaeoecology*, 196, 209-228.

891

892 Myrow, P.M., Southard, J.B., 1996. Tempestite deposition. *Journal of Sedimentary Research*, 66
893 (5), 875-887.

894

895 Neale, J.W., 1988. Ostracods and palaeosalinity reconstruction. In: De Deckker, P. Colin, J.-P.,
896 Peypouquet, J.P., (Eds), *Ostracoda in the Earth Science*, Elsevier, Amsterdam, pp. 125–155.

897

- 898 Nilsen, T. H., 1982. Fluvial Models. In: Sandstone Depositional Environments, Ed. P.A. Scholl &
899 D. Spearing. pp115-137. AAPG Memoir 31.
900
- 901 Orton, G.J., 1988. A spectrum of Middle Ordovician fan deltas and braidplain deltas, North Wales:
902 a consequence of varying fluvial clastic input. In: Nemeč, W., Steel, R.J. (Eds) Fan Deltas:
903 Sedimentology and Tectonic Settings. Blackie, London, 23-49.
904
- 905 Över, S., Ünlügenç, U. C., Bellier, O., 2002. Quaternary stress regime change in the Hatay region,
906 SE Turkey. Geophysics Journal International, 148, 649-662.
907
- 908 Parlak, O., Kop, A., Ünlügenç, U. C., Demirkol, C., 1998. Geochronology and geochemistry of
909 basaltic rocks in the Karasu graben around Kirikhan Hatay, S. Turkey. Turkish Journal of Earth
910 Sciences, 7, 53-61.
911
- 912 Patton, D.K., 1993. Samgori field, Republic of Georgia: critical review of island-arc oil and gas.
913 Journal of Petroleum Geology, 16, 153-167.
914
- 915 Piskin, O., Delaloye, M., Selçuk, H., Wagner, J., 1986. Guide to Hatay Geology SE Turkey.
916 Oflioliti, 11, 87-104.
917
- 918 Postma, G., 1986. Classification for sediment gravity-flow deposits based on flow conditions during
919 sedimentation. Geology, 14, 291-294.
920
- 921 Postma, G., 1990. Depositional architecture and facies of rivers and fan deltas: a synthesis. In:
922 Collela, A., Prior, D.B., (Eds) Coarse-grained deltas. Special Publication of the International
923 Association of Sedimentologists, 10, 13-27.

924

925 Reading, H. G. & Collinson, J. D. 1996. Clastic coasts. In: *Sedimentary Environments: Processes,*
926 *Facies and Stratigraphy*, 3rd edition Ed: H.G. Reading. Blackwell Science. 154-231.

927

928 Robertson, A. H. F., 1998. Mesozoic-Tertiary Tectonic evolution of the easternmost Mediterranean
929 area: integration of marine and land evidence. In: Robertson, A. H. F., Emeis, K.-C., Richter, C.,
930 Camerlenghi, A. (eds.), *Proceedings of the Ocean Drilling Program, Scientific Results*, 160, 723-
931 782.

932

933 Robertson, A.H.F., 2000. Mesozoic-Tertiary tectonic-sedimentary evolution of a south Tethyan
934 oceanic basin and its margins in southern Turkey. In: Bozkurt, E., Winchester, J.A, Piper., J.D.
935 (eds)., *Tectonics and Magmatism in Turkey and the Surrounding Area.*. Geological Society,
936 London, Special Publications, 173, 43-82.

937

938 Robertson, A.H.F., 2002. Overview of the genesis and emplacement of Mesozoic ophiolites in the
939 Eastern Mediterranean Tethyan region, *Lithos*, 65, 1-67.

940

941 Robertson, A. H. F, Dixon, J. E., 1984. Introduction: aspects of the geological evolution of the
942 Eastern Mediterranean. In: Dixon, J.E., Robertson, A. H. F (eds). *The Geological evolution of the*
943 *Eastern Mediterranean.* Geological Society , London, Special Publications, 17, 1-74.

944

945 Robertson, A. H. F., Ünlügenç, U. C., Inan, N., Tasli, K. 2004. The Misis-Andirin Complex: a
946 Mid-Tertiary melange related to late-stage subduction of the Southern Neotethys in S Turkey.
947 *Journal of Asian Earth Sciences*, 22, 413-453.

948

- 949 Robertson, A. H. F., Ustaömer, T., Parlak, O., Ünlügenç, U. C., Taşlı, K., İnan, N., 2006. The Berit
950 transect of the Tauride thrust belt, S Turkey: Late Cretaceous – Early Cenozoic
951 accretionary/collisional processes related to closure of the Southern Neotethys. *Journal of Asian*
952 *Earth Sciences*, 27, 108-145.
- 953
- 954 Rojay, B., Heimann, A., Toprak, V., 2001. Neotectonic and volcanic characteristics of the Karasu
955 fault zone Anatolia, Turkey: The transition zone between the Dead Sea transform and the East
956 Anatolian fault zone. *Geodynamica Acta*, 14, 197-212.
- 957
- 958 Rosenfeld, A., Hirsch, F., 2005. The Paleocene – Eocene of Israel. In: Hall, J. K., Krasheninnikov,
959 V. A., Hirsch, F., Benjamini, C., Flexer, A. (eds). *Geological Framework of the Levant, Volume II:*
960 *The Levantine Basin and Israel. Historical Productions – Hall, Jerusalem.* 437-459.
- 961
- 962 Saller, A., Armin, R., Ichram, L.O., Sullivan, C., 1993. Sequence stratigraphy of aggrading and
963 backstepping carbonate shelves, Oligocene, Central Kalmantan, Indonesia. In: Loucks, R. G., Sarg,
964 J. F. (eds), *Carbonate Sequence Stratigraphy: Recent developments and Applications.* American
965 *Association of Petroleum Geologists Memoir*, 57, 267-290.
- 966
- 967 Şaroğlu, F., Emre Ö., & Kuşçu, İ. 1992. The East Anatolian fault zone of Turkey. *Annales*
968 *Tectonicae*, 6, 99-125.
- 969
- 970 Şengör, A.M.C., Yılmaz, Y., 1981. Tethyan evolution of Turkey: a plate tectonic approach.
971 *Tectonophysics*, 75, 81-241.
- 972
- 973 Schwan, W., 1971. Geology and tectonics of the central Amanos Mountains. In: Cambell, A.S (ed)
974 *Geology and History of Turkey. Tripoli: The petroleum Exploration Society of Libya*, 283-303.

975

976 Sharland, P.R., Casey, D.M., Davies, R.B., Simmons, M.D., Sutcliffe, O.E., 2004. Arabian Plate
977 stratigraphy-revisions to SP2. *GeoArabia*, 9, 199-214.

978

979 Tatar, O., Piper, J. D. A., Gürsoy, H., Heimann, A., Koçbulut, F., 2004. Neotectonic deformation in
980 the transition zone between the Dead Sea Transform and the East Anatolian Fault Zone, Southern
981 Turkey: a palaeomagnetic study of the Karasu Rift volcanism. *Tectonophysics*, 385, 17-43.

982

983 Vincent, S. J., Allen, M. B., Ismail-Zadeh, A. D., Flecker, R., Foland, K. A., Simmons, M. D.,
984 2005. Insights from the Talysh of Azerbaijan into the Paleogene evolution of the South Caspian
985 region. *Geological Society of America Bulletin*, 11, 1513 – 1533.

986

987 Vincent, S. J., Morton, A. C., Carter, A., Gibbs, S., Barabadze, T. G., 2007. Oligocene uplift of the
988 Western Greater Caucasus: an effect of initial Arabia-Eurasia collision. *Terra Nova*, 19, 160-166.

989

990 Westaway, R., 2003. Kinematics of the Middle East and Eastern Mediterranean updated. *Turkish
991 Journal of Earth Sciences*, 12, 5-46.

992

993 Westaway, R., Arger, J., 2001. Kinematics of the Malatya-Ovacik fault zone. *Geodynamica Acta*,
994 14, 103-131.

995

996 Wright V. P & Burchette, T. P. 1996. Shallow water carbonate environments. In: *Sedimentary
997 Environments: Processes, Facies and Stratigraphy*, 3rd edition Ed: H.G. Reading. Pp. 325-394.
998 Blackwell Science.

999

- 1000 Yılmaz, Y., Yiğitbaş, E., Genç, S. C., 1993. Ophiolitic and metamorphic assemblages of southeast
1001 Anatolia and the significance in the geological evolution of the orogenic belt. *Tectonics*, 12, 1280-
1002 1297.
- 1003
- 1004 Yurtmen, S., Guillou, H., Westaway, R., Rowbotham, G., Tatar, O., 2002. Rate of strike-slip
1005 motion on the Amanos Fault (Karasu Valley, southern Turkey) constrained by K-Ar dating and
1006 geochemical analysis of Quaternary basalts. *Tectonophysics*, 344, 207-246.
- 1007
- 1008 Zachos, J. C., Pagani, M., Sloan, E. T., Billups, K., 2001. Trends, rhythms, and aberrations in global
1009 climate 65 Ma to present. *Science*, 292, 686-694.
- 1010

1011

1012 **Tables**

Name	Age	Lithology	Boundaries	Thickness
Gökdere Fm	Tortonian - Messinian	Marl and litharenite	Conformable with Kepez Fm or unconformable on or faulted against older units. Upper boundary to younger sediments not exposed.	400 – 700m
Kepez Fm	Langhian	Wackestone and packstone	Lower boundary unconformable on K1c1 or conformable with Gökdere, upper transitional to Gökdere Fm.	15m
K1c1 Fm	Aquitanian – Burdigalian	Conglomerates, litharenites and mudstones	Unconformable with units above and below. Also fault contacts with Gökdere Fm	100 - 150m
Hacıdağ Fm	Palaeocene - Eocene	Calcarenite, wackestone, packstone and rudstone	Base conformable with Cona and Esmisek Fms unconformable on the ophiolitic complex. Angular unconformity with K1c1 and Gökdere Fms. Faulted contacts with other formations numerous.	> 400m

1013 Table 1. Summary of the characteristics of the lithostratigraphic units discussed in the text.

	Serinyo	Kan1	Kan2	Kan3	Gok 1	Gok 2	Gok 3	Gok 4
	1							
Serpentinite	0	0	27	10	95	80	33	40
Chert	14	9	12	2	0	0	3	2
Algal lmst	73	0	0	0	0	0	0	0
Chert + lmst	13	20	26	21	0	0	0	0
Bioclastic lmst	0	5	12	13	0	0	0	0
Nummul. lmst	0	0	0	0	0	0	4	1
Undiff lmst	0	74	42	35	5	20	5	0
Total	100	108	119	81	100	100	45	43

1014

1015 Table 2. Clast counts from conglomerate horizons at Serinyol (Ser), Kanlidere (Kan) and Gökdere
1016 (Gok), location of sites shown on figures 6 and 9. The clast count was undertaken by drawing a 1x1
1017 m grid upon the outcrop, 10 cm intersections were marked and the clast at each intersection
1018 counted, giving the clast composition of ~ 100 clasts at each location.

Sample No	SB73A	SB71A	SB77A	SB94A	SB142A
Age	L.Mio	L. Mio	L. Mio	U.Mio	U. Mio
S.Calcite	20	31	37.5	28.5	19.5
Micrite	2	1.5	10.5	6.5	7
Qtz(m)	1	7.5	0	10.5	5.5
Qtz(p)	1.5	6.5	0	10	11.5
Ophiolite	51.5	7	48.5	13.5	11.5
bioclast	0	0	1	2	0
mica	0	0.5	0	2	0
carbonate	5.5	19	1.5	12.5	30
siliciclastic	14.5	17	0.5	8	6
feldspar	0	1	0	4	0.5
opaque	4	4	0	2.5	2
other	0	5	0.5	0	6.5
Total	100	100	100	100	100

1019

1020 Table 3. Point-counting results for sandstones. Two hundred points were counted for each thin-
1021 section in order to have a statistically meaningful sample group. For locations of samples (SB73A,
1022 etc.) see Fig. 1. S.Calcite = sparry calcite; Qtz (m) = monocrystalline quartz; Qtz (p) =
1023 polycrystalline quartz.

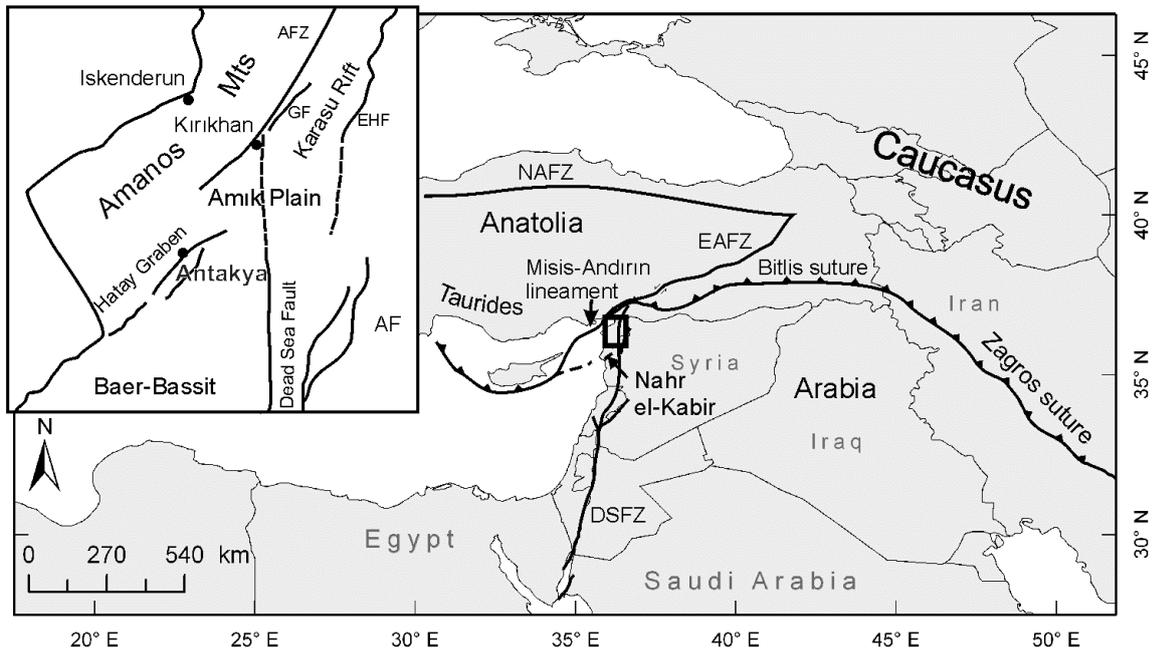
Sample	Age	Fm.	Quartz	Calcite	Smectite	Clinochrysotile	Albite	Dolomite	Muscovite	Others	Total
SB72A	U.Mio	G	29	51	0	0	12	2	0	6	100
SB95A	U.Mio	G	28	49	0	0	0	2	5	16	100
SB98A	U.Mio	G	14	57	0	8	2	0	16	3	100
SB140A	U.Mio	G	15	56	0.1	22	5	0	0	1.9	100
SB135A	U.Mio	G	38	59	0	3	0	0	0	0	100
SB137A	U.Mio	G	26	57	0	0	10	0	6	1	100
SB104A	L.Mio	K	32	41	0	0	10	0	0	17	100
SB76A	L.Mio	K	20	5	16	39	0	0	20	0	100
SB126A	Eocene	H	17	61	0	1	0	0	14	7	100
SB128A	Eocene	H	20	78	0	0	2	0	0	0	100
SB133A	Eocene	H	19	79	0	0	2	0	0	0	100

1024

1025 Table 4. XRD determinations of fine-grained sediments (sample locations shown on Fig. 1). Two
1026 samples were analysed from the K1c1 Formation (SB76A & SB104A), both have compositions rich
1027 quartz and clinochrysotile. Sample SB76A contains abundant smectite (16%) and muscovite (20%)
1028 but low calcite (5%). Sample SB104A, by comparison, has no smectite or muscovite but has a high
1029 (56%) calcite content. Six samples were analysed from the Gökdere Formation. All have
1030 significant amounts of quartz (14-38%) and high calcite (31-78%) contents. Albite and muscovite
1031 also are relatively abundant. Formation codes; G = Gökdere; K = K1c1 and H = Hacıdağ.

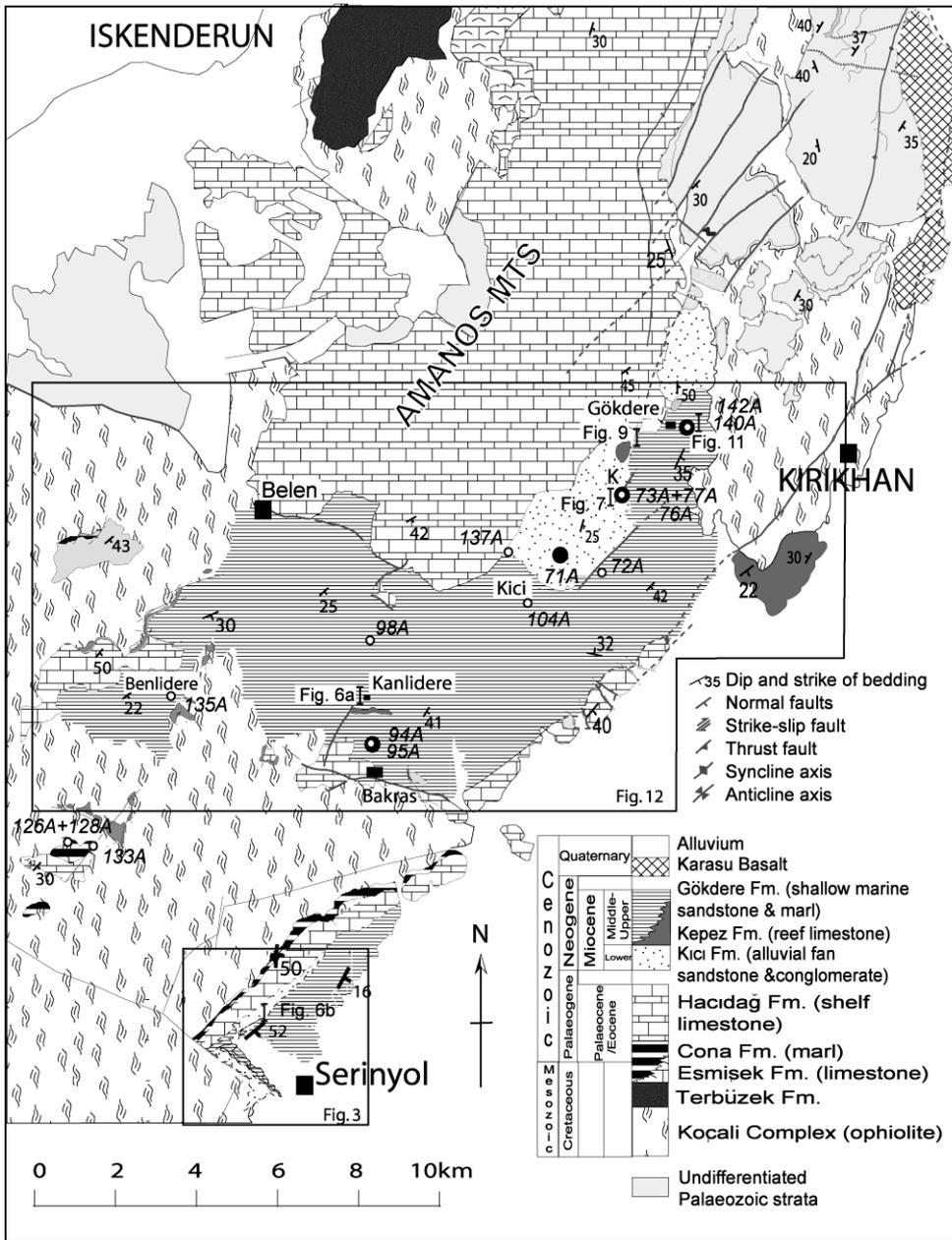
1032

1033 **Figures**



1034

1035 Figure 1. Regional geodynamic framework, small box shows location of the study area. Inset box
 1036 shows major faults of the study area and locale, DSFZ: Dead Sea Fault Zone; EAFZ, East Anatolian
 1037 Fault Zone; NAFZ, North Anatolian Fault Zone; AFZ, Amanos Fault Zone, EHF, East Hatay Fault;
 1038 AF, Afrin Fault; GF, Guzelce Fault.



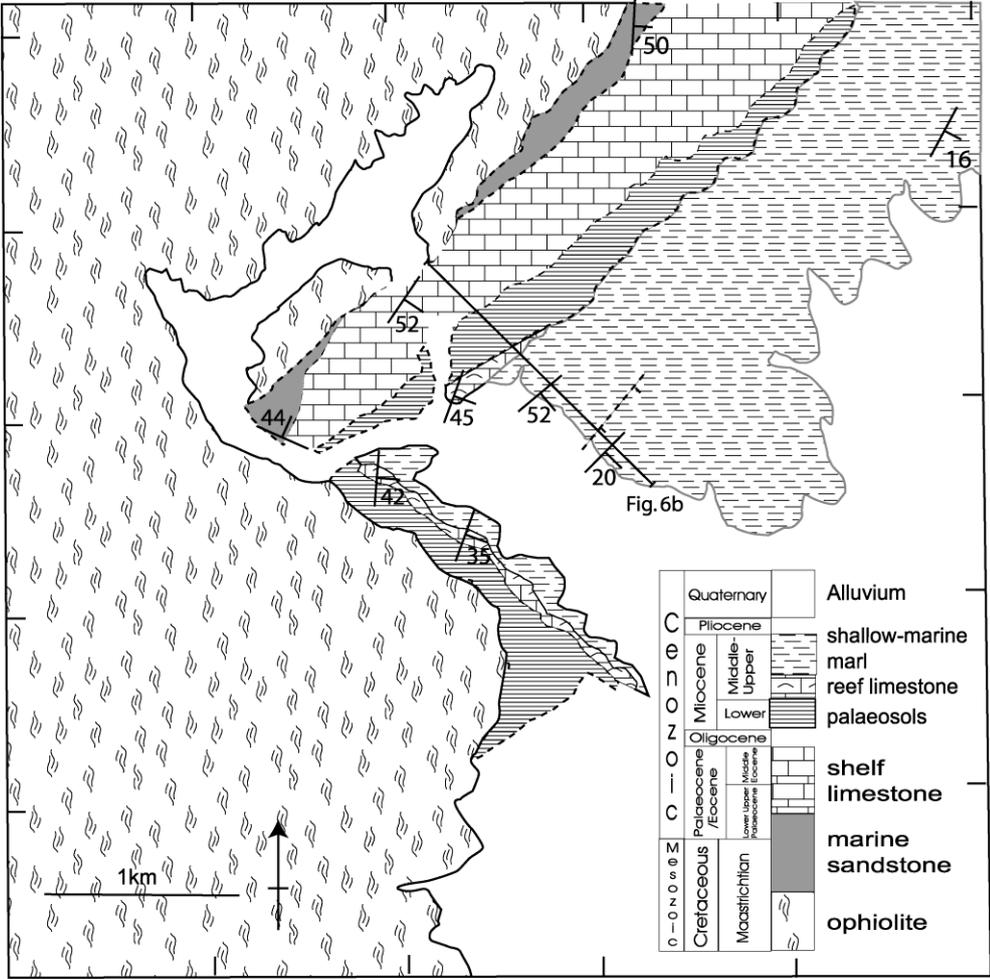
1039

1040

1041 Figure 2. Geological map of the northern part of the study area around the towns of Kırıkhan and
 1042 Belen, showing the main places discussed in the text and the locations of samples used in
 1043 petrological analysis (modified from Boulton et al., 2007); black circles indicate locations where
 1044 sandstones for point-counting were collected, white circles indicate locations where fine-grained
 1045 sediments were taken for XRD analysis, bars with appended Figure numbers indicate locations of
 1046 logs, while boxes indicate the extent of figures 3 and 13. The letter K indicates the location of the
 1047 type section of the Kıcı Formation at Kurtisoğuksu.

1048

1049



1050

1051

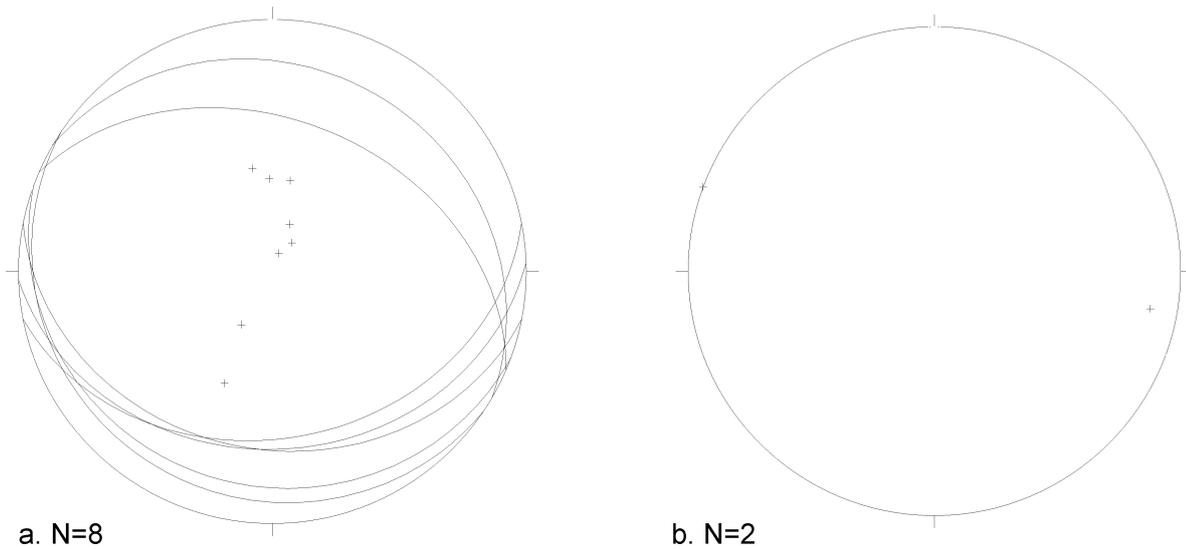
1052 Figure 3. Geological map of the area around Serinyol, see figure 2 for the location within the
 1053 Karasu Rift. Black line indicates the location of the log shown in figure 6b. Note the town of
 1054 Serinyol is located under the key.



1055

1056 Figure 4a. Incipient slump in thin-bedded Eocene limestone at 0255391/4036316 (Map sheet
 1057 Antakya P36-a2), north of the Kırıkhan-Antakya road; b). Isoclinal recumbent slump fold observed
 1058 in Eocene limestones (circled compass-clinometer provides scale) just off the main Belen road at
 1059 grid ref. 0252481/4041026 (Map sheet Antakya P36-a2). Orientations of this fold hinge and the
 1060 incipient slump shown in (a) demonstrate that the palaeoslope was inclined to the north (see fig. 5 for
 1061 bedding data); c). General view of the Eocene sediments near Serinyol at 0248123/4029755 (Map
 1062 sheet Antakya P36-a3). Three units can be observed the basal unit is ophiolite (Op), an erosive
 1063 unconformity separates the basement from the overlying Eocene composed of a basal conglomerate
 1064 (Con) and upper bedded limestones (Lm) above; d). View of the Eocene (Ha - RHS) – Lower
 1065 Miocene (K1 - LHS) boundary near the village of Gökdere, along which a stream has flowed.
 1066 Eocene limestones are folded, while the overlying K1c1 formation is unfolded, dipping
 1067 southeastwards (towards the camera). The Lower Miocene strata is formed of a basal algal
 1068 boundstone representing a brief marine incursion, with continental clastics above.

1069



1070
1071

1072

1073

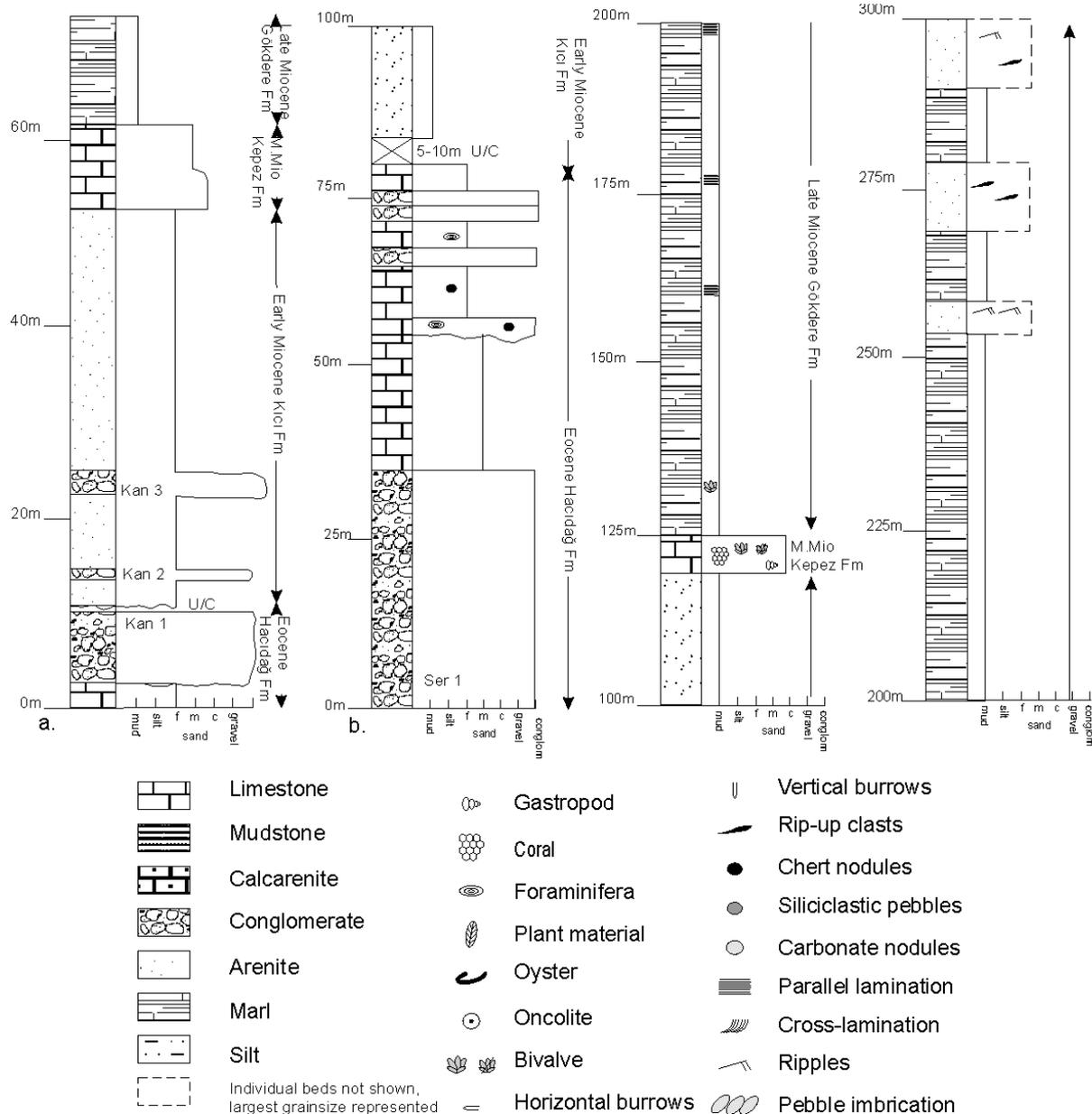
1074

1075

a. N=8
b. N=2

Figure 5a). Bedding plane data (shown as great circles and pole to plane - the crosses) for limbs of slump folds observed in the Eocene Hacıdağ Formation; b). Crosses indicate orientation of hinge lines of measured slump folds (Fig. 4b). Both sets of data indicate the general orientation of the palaeoslope was N to NNE; perpendicular to the hinges of the folds that form parallel to the strike of the palaeoslope.

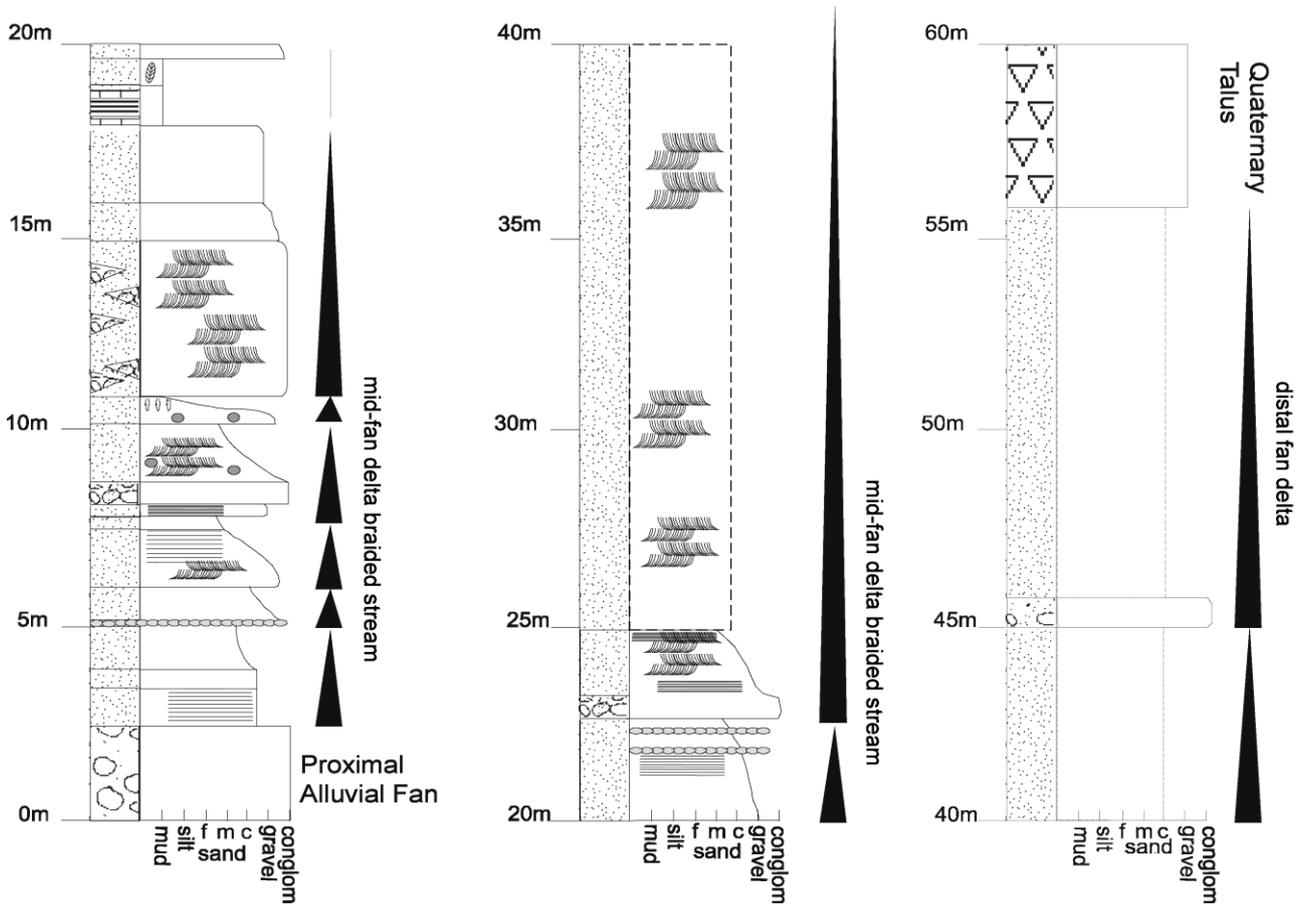
1076



1077

1078 Figure 6a). Log of Sediments at Kanlidere (on figure 2). Kan1, 2, 3 and Ser 1 show the positions
 1079 where conglomerate clast counts were undertaken, see Table 2 for results; b. Log of sediments at
 1080 Serinyol (see figure 3 for position of logged section). Key for all logs shown at the bottom.

1081



1082

1083

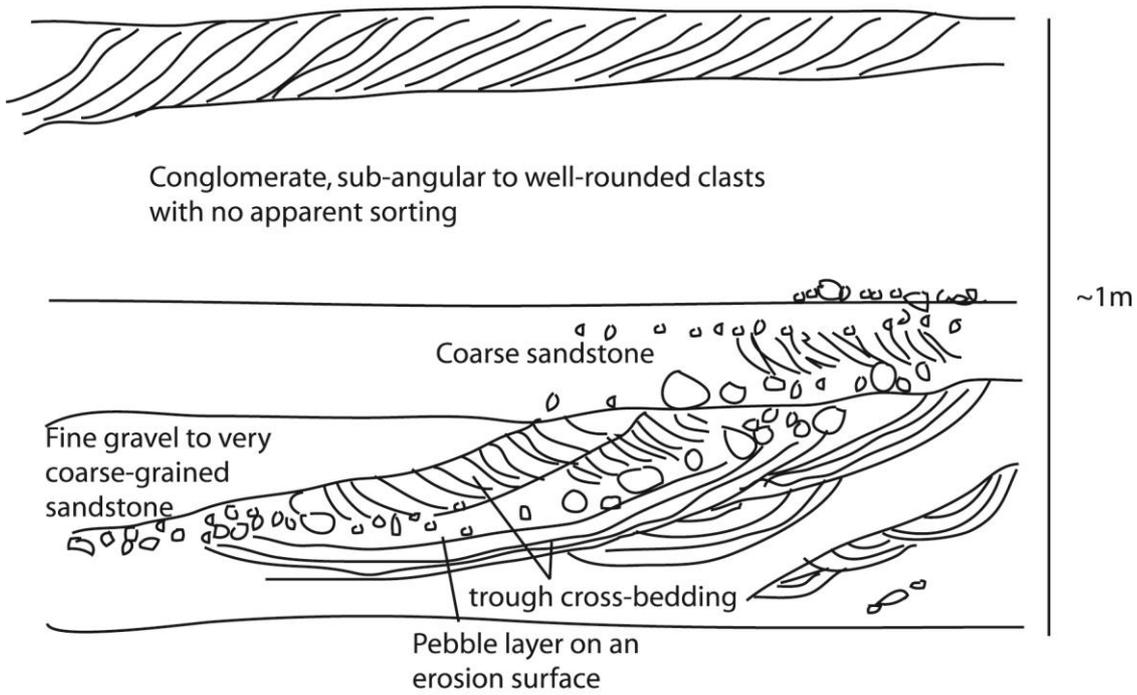
1084 Figure 7. Log of K1c1 formation at the type section (K; Fig. 2) with sedimentary environments and
 1085 fining-upwards cycles shown. Fining-upwards sequences may represent bar processes or where a
 1086 basal conglomerate is present, may be a channel fill deposit (modified from Boulton et al., 2007).

1087 See figure 6 for key.



NW

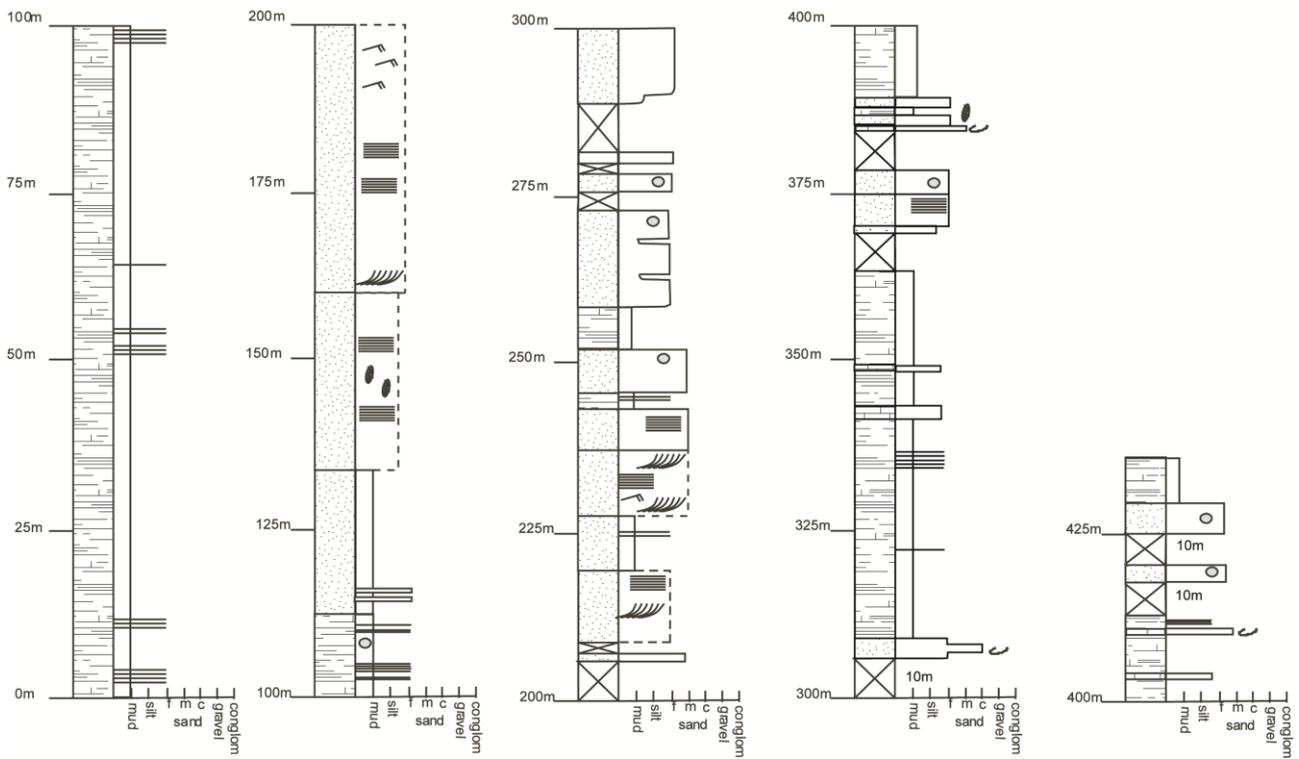
SE



1088

1089 Figure 8. Field photograph and sketch of cross-bedded sandstones typical of the K_{1c1} Formation,

1090 from the type section (K; Fig. 2). See figure 6 for key.



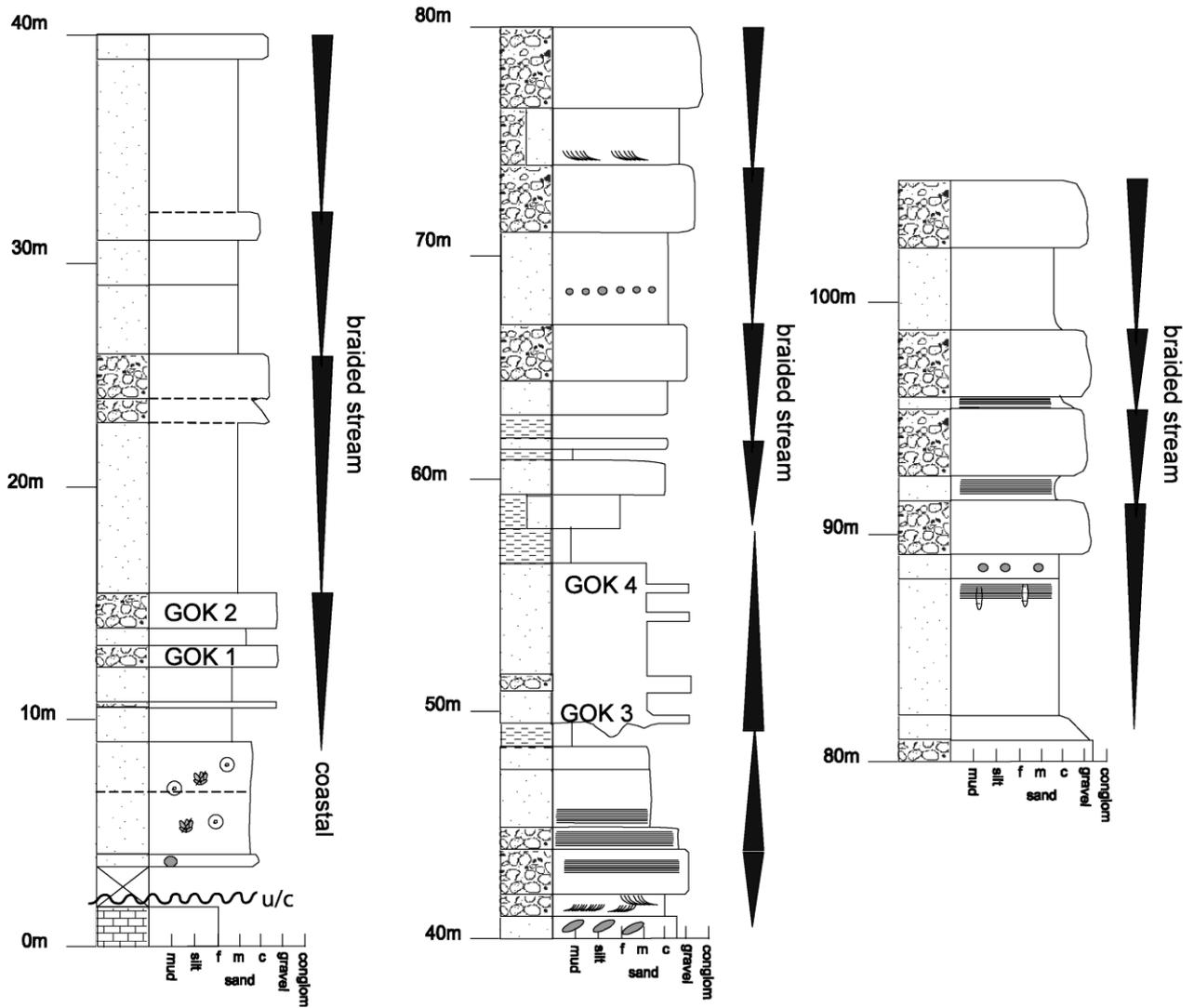
1091

1092 Figure 9. Log of the Kıcı Formation near the village of Gökdere (Fig. 2), the sequence is interpreted
 1093 as an alluvial fan deposit where coarsening-upwards cycles possibly represent progradation of
 1094 individual fan lobes. Gok 1, 2, 3, 4 refer to locations where conglomerate clast counts were
 1095 undertaken, results shown in table 2.



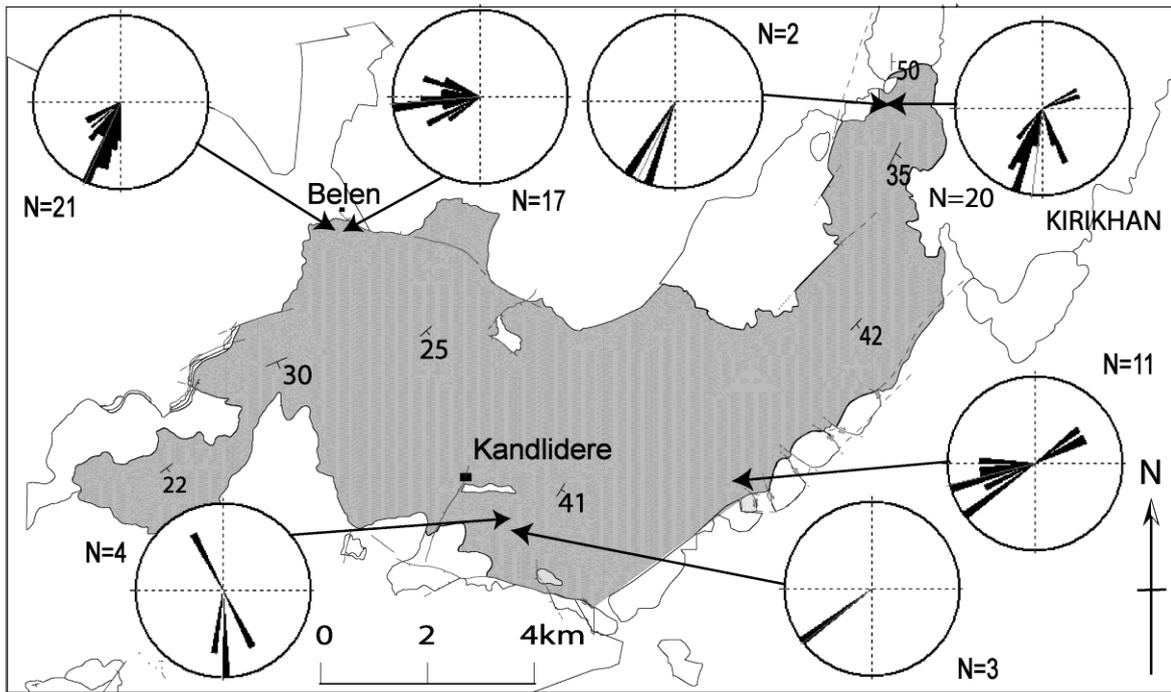
1096

1097 Figure 10a. Close up of a conglomerate horizon in the K1c1 Formation, at Kanlidere, note borings in
 1098 limestone clasts; b). Close-up of blocks of coralline limestone from the Kepez Formation observed
 1099 near Serinyol; c). View over the village of Gökdere with part of Cenozoic sedimentary sequence
 1100 exposed from left to right in the foreground. In the lower left the ophiolite complex (Op) is
 1101 observed, above is the Upper Miocene Gökdere Formation (Go) dipping southeastwards (to the
 1102 right). The Middle Miocene Kepez Formation (Ke) can be observed as a laterally discontinuous
 1103 outcrop exposed in the side of hill (in the centre of the field of view), the K1c1 Formation is not
 1104 present at this location; d). View of the middle of the Gökdere Formation at the type section
 1105 showing a typical sand-body on the left hand side.



1106

1107 Figure 11. Log of the Gökdere Formation at Gökdere (Fig. 2). See figure 6 for key. The sequence is
 1108 interpreted as a regressive shallow-marine sequence.

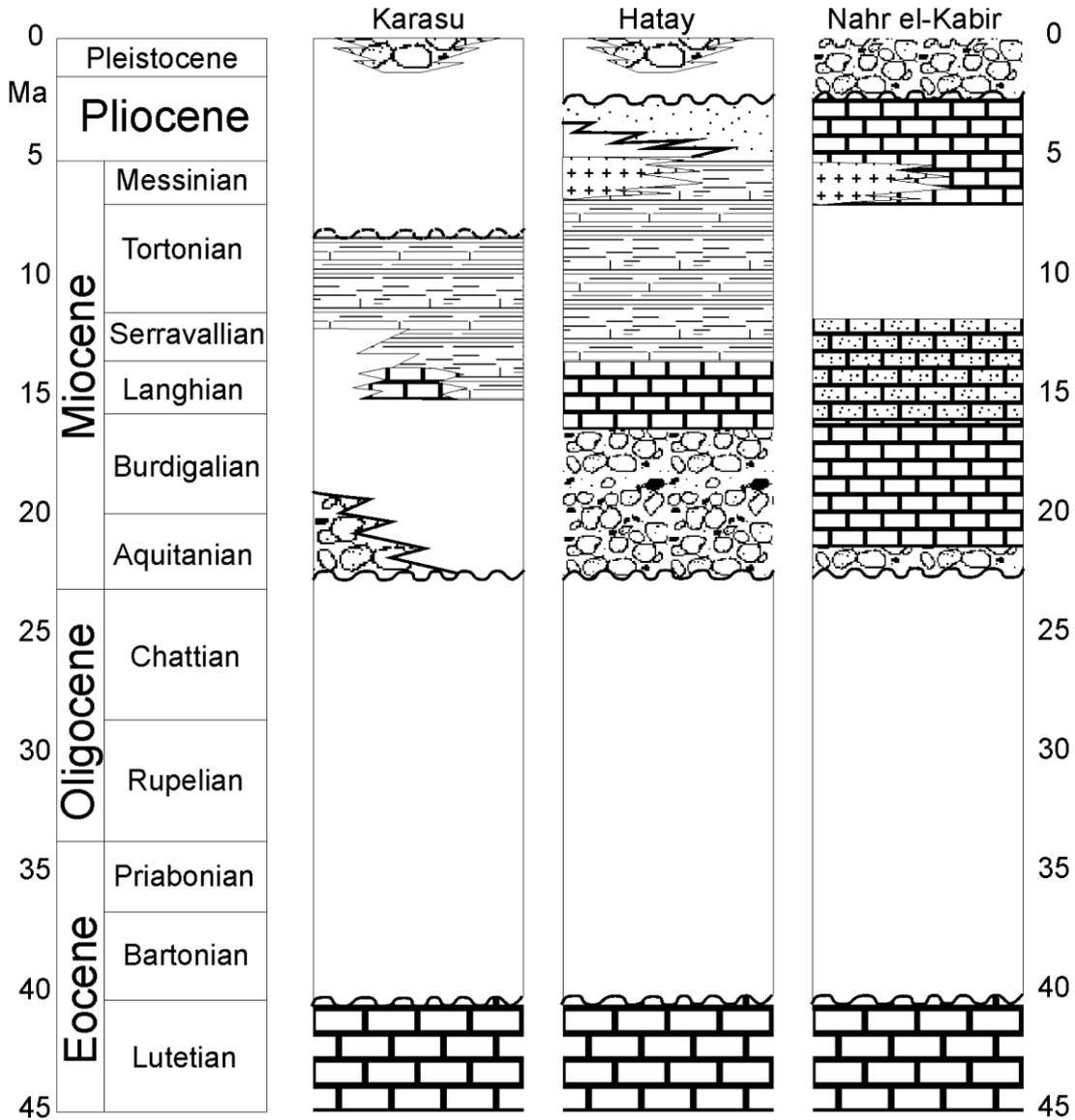


1109

1110 Figure 12. Rose diagrams of measured palaeocurrents from the Upper Miocene Gökdere Formation
 1111 in the central part of the study area (extent of this map shown on figure 2). N = No. of measurements
 1112 at each locality.

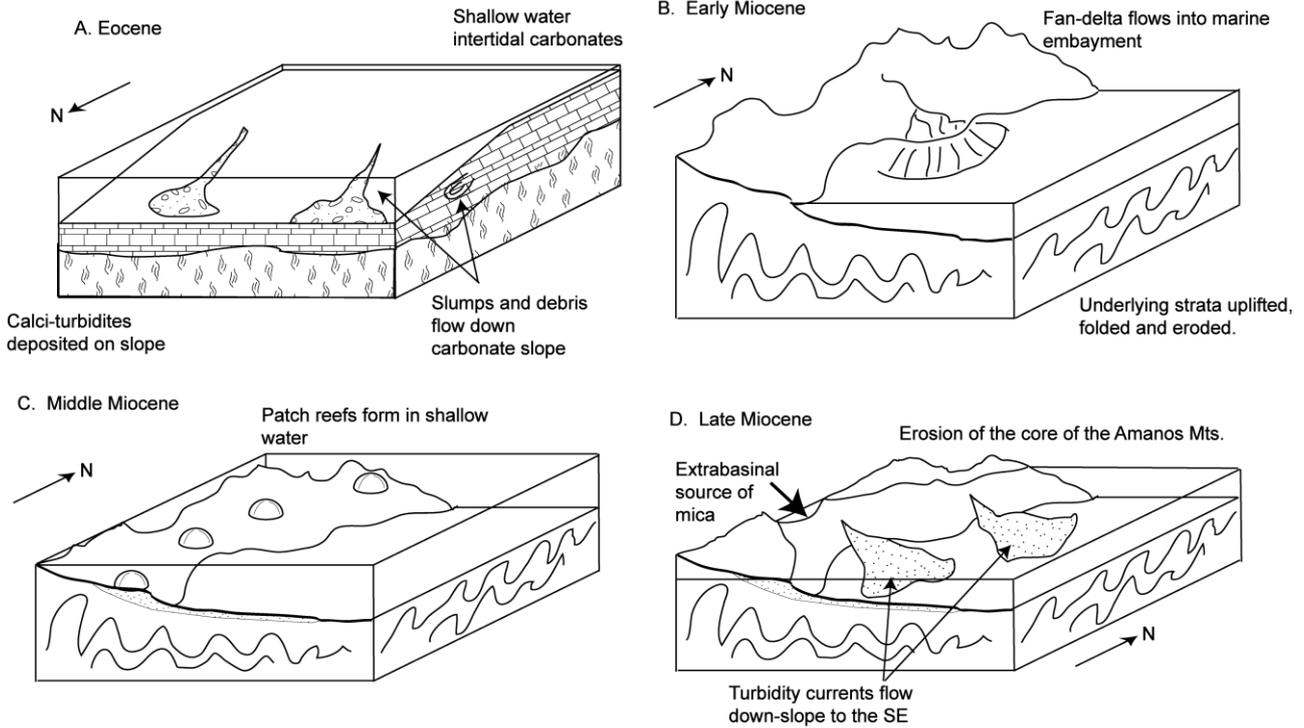
1113

1114



1115

1116 Figure 13. Stratigraphic columns showing the stratigraphy of the Karasu Rift, Hatay Graben and
 1117 Nahr el-Kabir Graben (locations of areas shown on Figure 1). Note the similarities between the
 1118 Karasu Rift and adjacent Hatay Graben and the differences of both areas to the Nahr el-Kabir
 1119 Graben to the south. See figure 6 for key.



1120

1121 Figure 14. Block diagrams illustrating inferred palaeogeographical conditions during A) Eocene; B)

1122 Lower Miocene; C) Middle Miocene; D) Upper Miocene and E) Pliocene to Recent times. N.B.

1123 Note the reversed orientation of diagram (A).

1124

1125

1126

1127

1128

1129

1130