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Validating far-field deformation styles from the Adjara-Trialeti fold-and-thrust belt to the Greater Caucasus (Georgia) through multi-proxy thermal maturity datasets

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1 **Validating far-field deformation styles from the Adjara-Trialeti fold-and-thrust belt to the**
2 **Greater Caucasus (Georgia) through multi-proxy thermal maturity datasets**

3

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17

18

19 **Abstract**

20

21 Thermal history reconstructions can help to better characterise the geological history of areas
22 that experienced a polyphase tectonic evolution. The integration of published
23 stratigraphic/structural data with new and pre-existing data on thermal maturity (clay mineralogy,
24 Raman spectroscopy, vitrinite reflectance, and pyrolysis) of both surface and subsurface
25 sedimentary successions of a wide region of Georgia including -north to south- the southern Greater

26 Caucasus, the western Kura Basin, and the Adjara-Trialeti fold-and-thrust belt (FTB) provides
27 cogent constraints on its late Mesozoic-Cenozoic tectono-sedimentary evolution.

28 Overall, thermal maturity spans from the low diagenesis (60-80°C) in the Upper Miocene section
29 of the Kura Basin to anchizone-epizone (about 400°C) in the central Greater Caucasus axial zone.
30 In more detail, different maturity trends and thermal histories point to the existence of two domains
31 formed by positive tectonic inversion: (i) the Adjara-Trialeti FTB from an Eocene rift basin and (ii)
32 the Greater Caucasus from a Mesozoic rift basin. Multiple thermal indicators, along with
33 stratigraphic/structural evidence, show that the Paleocene section of the Adjara-Trialeti basin fill
34 reached the upper oil window (ca. 115°C) during maximum sedimentary burial and that the whole
35 basin was then exhumed starting from the late Middle Miocene. A positive correlation between
36 thermal maturity and stratigraphic age points to a limited thermal effect of tectonic loading. In the
37 southern Greater Caucasus, thermal maturity increases progressively with stratigraphic age, from
38 ca. 100°C (Upper Eocene) to 400°C (Lower Jurassic), in broad agreement with the reconstructed
39 thickness of the basin-fill succession, thus indicating that most of the thermal maturity was again
40 induced by sedimentary burial.

41 As to the flexural western Kura Basin, its Maikopian (Oligocene-Early Miocene) section reached
42 the oil window (up to ca. 110°C) whereas the Middle-Late Miocene one is immature. The Kakheti
43 ridge -a highly tectonised portion of the Kura Basin- reached immature to early mature conditions.

44

45 **Keywords**

46 *Intra-continental deformation, Alpine orogeny, Maikop, thermal indicators, Caucasus, Kura Basin*

47

48 **1. Introduction**

49

50 The use of indicators of maximum paleo-temperatures and thermal maturity from sedimentary
51 successions in orogenic zones is traditionally used for hydrocarbon (HC) exploration (e.g. Aldega et
52 al., 2014; Allen and Allen, 2013; Tozer et al., 2020). Less frequently, it is applied to validate
53 structural styles in deformed orogenic belts (e.g. Aldega et al., 2018; Atouabat et al., 2020; Balestra
54 et al., 2019; Caricchi et al., 2015; Di Paolo et al., 2014; Muirhead et al., 2020; Tozer et al., 2020),
55 either because of lack of constraints on timing of exhumation that can bias thermal modelling, or
56 because such indicators mostly derive from surface outcrops and can allow modelling only of
57 pseudo-well sections, rather than present-day boreholes, introducing an extra degree of uncertainty.

58 In recent years the frequent integration of classical and cutting-edge indicators of thermal
59 maturity due to burial (either sedimentary or tectonic) allowed the assessment of maximum paleo-
60 temperatures in sedimentary basins with reduced error bars (Corrado et al., 2005, 2020; Goodhue
61 and Clayton, 2010; Labeur et al., 2021; Liu et al., 2019; Mangenot et al., 2017, 2019; Qiu et al.,
62 2020; Spina et al., 2018). For example, pyrolysis parameters (HI, PI, Tmax) (Behar et al., 2001;
63 Tissot et al., 1987), clay-derived geothermometers (such as illite percentage in illite-smectite mixed
64 layers and illite crystallinity index, KI; Aldega et al., 2007a, 2007b; Schito et al., 2016), vitrinite
65 reflectance (Balestra et al., 2019; Burnham and Sweeney, 1989; Corrado et al., 2009; Dow, 1977) in
66 the diagenetic realm and Raman spectroscopy parameters in both the metamorphic and diagenetic
67 realms on organic matter (Beyssac et al., 2002; Lahfid et al., 2010; Lünsdorf and Lünsdorf, 2016;
68 Schito et al., 2017) can lead to significant reduction of admissible paleotemperature ranges in the
69 evolution of compressional areas, especially when they are combined with maximum
70 paleotemperatures derived from low-T thermochronological modelling [fission-track and (U-
71 Th)/He dating on apatite crystals] (Aldega et al., 2011; Corrado et al., 2020; Schito et al., 2018).

72 Georgia, located in the deformed hinterland of the Arabia-Eurasia collision occurring along the
73 Bitlis-Zagros suture zone, represents a privileged and fascinating natural laboratory to validate
74 structural styles that developed during Arabia-Eurasia convergence using thermal maturity datasets.
75 Here, different orogenic chains crop out with opposite vergences, variable structural styles and

76 shortening degrees, accommodating far-field regional convergence (e.g. Adamia et al., 2010,
77 2011b; Alania et al., 2017; Nemčok et al., 2013). In the present study we consider three tectonic
78 domains: from south to north they are (i) the Adjara-Trialeti FTB, (ii) the Kura Basin (comprising
79 its northern highly deformed portion, the Kakheta ridge) and (iii) the Georgian Greater Caucasus
80 (Fig. 1). Brittle structures and basin sedimentary fills, due to stretching developed either in
81 Mesozoic or early Cenozoic times, influence the geometry and distribution of the late Cenozoic
82 compressive deformation that brought to minor (Kura Basin and Kakheta ridge), moderate (Adjara-
83 Trialeti FTB) or intense exhumation (Georgian Greater Caucasus), with a peak in Miocene times
84 (Alania et al., 2017; Gusmeo et al., 2021; Vincent et al., 2020).

85 In this region, geometric and genetic relationships between areas affected by positive inversion
86 and moderate to high exhumation, and areas where thin-skinned thrust tectonics develops with
87 higher shortening and less exhumation, are not consistently described (Adamia et al., 2010; Alania
88 et al., 2017, 2018, 2020; Mosar et al., 2010; Nemčok et al., 2013). Different seismic interpretations
89 and scarcity of detailed structural surveys have led to contrasting structural interpretations (Adamia
90 et al., 2010, 2011b; Alania et al., 2020; Banks et al., 1997; Forte et al., 2010, 2013, 2014; Mosar et
91 al., 2010; Nemčok et al., 2013; Tibaldi et al., 2017, 2018). There is also uncertainty regarding the
92 eastward continuation of the Adjara-Trialeti FTB in easternmost Georgia and Azerbaijan, and its
93 link with the unconstrained retrowedge of the Lesser Caucasus (Alania et al., 2017; Nemčok et al.,
94 2013; Sosson et al., 2010, 2016). Moreover, the geometry of the main structural features within the
95 Greater Caucasus, at least in Georgia, is only shown at a crustal scale (e.g. Mosar et al., 2010;
96 Nemčok et al., 2013; Saintot et al., 2006; Sosson et al., 2016). General agreement exists on the
97 nature of a thin-skinned south-verging thrust system in the western portion of the Kura Basin to the
98 south of the eastern Greater Caucasus, where HC exploration and production are ongoing (Alania et
99 al., 2017, 2018; Pace et al., 2019; Pupp et al., 2018).

100 The purpose of this paper is to give new constraints on the structural style in the three domains
101 of continental deformation in Georgia, to the north of the Bitlis-Zagros suture zone, by presenting
102 and integrating two thermal maturity datasets:

103 • Surface data, derived from clay mineralogy and Raman spectroscopy, petrography and
104 pyrolysis on organic matter, from Jurassic to Upper Miocene lithostratigraphic units. Original data
105 generated during this study have been integrated by published data from the Kura Basin and the
106 Greater Caucasus.

107 • Subsurface data, including both published and unpublished results from deep wells
108 exploring the Oligocene-Lower Miocene Maikop series, in the western portion of the Kura Basin
109 and in the easternmost Adjara-Trialeti FTB. These data result from the prolonged attention devoted
110 to the Maikop series, recognised as the main source rock in the Kura Basin (Boote et al., 2018;
111 Pupp et al., 2018). Its oil potential is quite low because of low TOC values and prevalence of type
112 III kerogen (rich in terrestrial input) dispersed in sedimentary rocks. These features extend also to
113 the east moving towards the Caspian Sea (Washburn et al., 2019). Nevertheless, the significant
114 burial depth within the Kura Basin allowed source intervals to enter the oil window in the
115 surroundings of Tbilisi (Pupp et al., 2018; Sachsenhofer et al., 2018).

116 The integration of the two datasets with the tectono-stratigraphic setting of the three main
117 structural domains, derived from original field surveys and pre-existing literature, allowed (i) to
118 constrain the level of thermal maturity acquired through time in the extensional and flexural basins
119 considered, and (ii) to evaluate the relative contribution of sedimentary/tectonic burial to the
120 thermal maturation during extensional phases and intraplate shortening, a few hundred kilometres to
121 the north of the Bitlis-Zagros suture zone of the Arabia-Eurasia collision (Cavazza et al., 2018,
122 2019).

123

124 2. Geological Setting

125

126 The study area is located in eastern Georgia and covers (i) the easternmost Adjara-Trialeti fold-
127 and-thrust belt, (ii) the southern (Georgian) side of the central Greater Caucasus orogen (GC), (iii)
128 the westernmost Kura Basin (including the Kakheti ridge, a structural culmination developed in its
129 northern sector) (Fig. 1).

130 The Adjara-Trialeti FTB is an orogen bordered mainly by north-vergent frontal reverse faults
131 (Alania et al., 2018; Banks et al., 1997; Gusmeo et al., 2021; Nemčok et al., 2013) and resulting
132 from the structural inversion of a former back-arc rift basin developed on the upper (Eurasian) plate
133 of the northern Neotethys subduction zone (Adamia et al., 1981, 2011b; Banks et al., 1997; Barrier
134 et al., 2018; Lordkipanidze et al., 1989). The main phase of rifting occurred in the Middle Eocene,
135 characterized by the deposition of a thick volcanic and volcanoclastic succession, accompanied by
136 shallow mafic-to-intermediate intrusions (Adamia et al., 2011b; Banks et al., 1997; Okrostsvardze
137 et al., 2018; Yılmaz et al., 2000, 2014). The post-rift phase lasted from the Late Eocene/Oligocene
138 to the Early Miocene and was followed by structural inversion since late Middle Miocene times
139 (Gusmeo et al., 2021). The Adjara-Trialeti FTB is often considered as part of the retro-wedge of the
140 Lesser Caucasus orogen *s.l.*, despite having an independent origin (Alania et al., 2017; Mosar et al.,
141 2010; Nemčok et al., 2013; Yılmaz et al., 2014).

142 The Greater Caucasus orogen is a fold-and-thrust belt resulting from the inversion of a back-arc
143 rift basin (Greater Caucasus Basin) which opened in the Early Jurassic (Adamia et al., 1981, 2011a,
144 2011b; Dercourt et al., 1986; Mosar et al., 2010; Nikishin et al., 2001; Saintot et al., 2006;
145 Sobornov, 1996; Zonenshain et al., 1990). Rifting is marked by Hettangian-Sinemurian black shales
146 nonconformably overlying the crystalline basement, followed by siliciclastic turbidites, lavas and
147 volcanoclastics deposited until the late Middle Jurassic; volcanic products were mostly deposited
148 from Aalenian to Bajocian times (Adamia et al., 2011a, 2011b; Lordkipanidze et al., 1989; Nikishin
149 et al., 2001; Saintot et al., 2006). From the latest Middle Jurassic until the Late Eocene the basin
150 experienced post-rift thermal subsidence, characterized by the deposition of calcareous and

151 siliciclastic turbidites (Adamia et al., 2011a, 2011b; Saintot et al., 2006; Zonenshain et al., 1990).
152 The Greater Caucasus Basin was probably underlain by thinned continental crust rather than
153 oceanic crust (Ershov et al., 2003).

154 There is an ongoing debate regarding the timing of structural inversion of the Greater Caucasus
155 Basin and subsequent development of the Greater Caucasus orogen, with hypotheses ranging from
156 the earliest Oligocene (Lozar and Polino, 1997; Nikishin et al., 2017; Vincent et al., 2007, 2013a,
157 2013b, 2016) to the Middle Miocene (Rolland, 2017) to the Pliocene (Avdeev and Niemi, 2011;
158 Cowgill et al., 2016; Forte et al., 2014; Philip et al., 1989). Low-temperature thermochronology
159 data seem to suggest an earlier growth of the western Greater Caucasus (e.g. Vincent et al., 2011)
160 with respect to the eastern and central parts of the orogen (e.g. Avdeev and Niemi, 2011; Vasey et
161 al., 2020; Vincent et al., 2020). The central and eastern portions of the Greater Caucasus certainly
162 underwent rapid Pliocene to recent uplift. There is no consensus on the causes of such a fast
163 exhumation (see for example the discussion in Vincent et al., 2020). Anyway, most authors agree
164 that at least about 5-8 km of Cenozoic uplift occurred in the Greater Caucasus.

165 Convergence between the Greater Caucasus and the Lesser Caucasus, namely the Adjara-Trialeti
166 FTB in the study area, caused the development of the so-called Transcaucasian intermontane
167 depression, constituted by the Kura and Rioni flexural foreland basins, plunging to the east and
168 west, respectively, and separated by the Dzirula Massif (Adamia et al., 2010, 2011b; Alania et al.,
169 2017; Banks et al., 1997; Nemčok et al., 2013; Rolland et al., 2011). The two basins developed as a
170 flexural response to both the Greater Caucasus to the north and the Lesser Caucasus *s.l.* to the
171 south, and are filled by Oligocene-to-recent sediments (Fig.2) (Adamia et al., 2010; Banks et al.,
172 1997; Nemčok et al., 2013). The Kartli Basin (Figs. 1 and 2) is considered as a sub-basin of the
173 Kura foreland basin bordered by the Adjara-Trialeti FTB to the south and the Greater Caucasus to
174 the north. In the Kura Basin, the Maikop series is composed of Oligocene-Lower Miocene clastic
175 (shales, siltstones and fine-grained sandstones) and evaporitic rocks deposited in the anoxic-dysoxic
176 environment of the Paratethys (Pupp et al., 2018; Sachsenhofer et al., 2018). The thickness of the

177 Maikop series can reach 2.5-3.5 km in some parts of the Kura Basin (Adamia et al., 2010). During
178 Middle to early Late Miocene times further 1.5-2.2 km of shales and fine-grained siliciclastics
179 (sandstones), intercalated in the uppermost sections with mainly calcareous units (mudstones, marls
180 and oolitic limestones and locally with coarse-grained rocks), were deposited within the Kura Basin
181 (Adamia et al., 2010; Alania et al., 2017). Since the Tortonian, the western part of the basin has
182 been under subaerial conditions and marine conditions persisted only in some localities and in the
183 easternmost portion of the basin (Adamia et al., 2010; Barrier et al., 2018). At the same time the
184 widespread deposition of coarse-grained clastic deposits, eroded from the adjacent orogenic belts,
185 started. Continental conditions prevailed from the Late Sarmatian (i.e. Tortonian, Fig. 2; see
186 Adamia et al., 2010; Lazarev et al., 2019; Neubauer et al., 2015 for a review) to the present,
187 interrupted only in the late Pliocene by a short-lived shallow marine environment, probably in
188 response to the rapid growth and advancement of the Greater Caucasus and the ensuing subsidence
189 in the foreland area (Adamia et al., 2010; Avdeev and Niemi, 2011; Nemčok et al., 2013;
190 Sukhishvili et al., 2020).

191 Nemčok et al. (2013), based on the geometry of the sedimentary wedges, recognized a multi-
192 stage development of the Kura foreland basin. In Oligocene times the depocenter was located along
193 the SW border with the sedimentary fill thinning progressively towards the NE. In Early-Middle
194 Miocene times maximum subsidence switched to the NE, with a clastic wedge progressively thinner
195 and finer grained from NE to SW, indicating that the basin was being flexed in response to the
196 southward advance of the Greater Caucasus. Since the Late Sarmatian (Tortonian) ongoing
197 convergence between the Greater and Lesser Caucasus forced the final uplift of the Dzirula Massif
198 and the Kura Basin started plunging towards the Caspian Sea, as demonstrated by the progressive
199 emergence of the basin from west to east (Adamia et al., 2010; Alania et al., 2017; Nemčok et al.,
200 2013). Symmetrically, the Rioni Basin plunged towards the Black Sea, on the western side of the
201 Dzirula Massif. Thus, the Dzirula Massif basement high separated definitely the Rioni Basin from
202 the Kura Basin (Banks et al., 1997; Barrier et al., 2018; Shatilova et al., 2020).

203 Continued convergence between the Greater and Lesser Caucasus caused incremental
204 deformation of the Kura foreland basin. Thick-skinned deformation occurred in the Early-Middle
205 Sarmatian (late Serravallian-early Tortonian) followed by thin-skinned deformation from the Late
206 Sarmatian-Meotian (Tortonian) onward (Nemčok et al., 2013). During convergence, the Greater
207 Caucasus deformation propagated into the northern Kura Basin forming the Kura south-vergent
208 thin-skinned foreland FTB (Kakheti ridge), starting from the Middle Miocene, with peak
209 deformation in the Late Miocene-Pliocene (Alania et al., 2017). A Late Pliocene-Pleistocene
210 acceleration of uplift occurred also in this belt (Sukhishvili et al., 2020), probably linked with
211 coeval enhanced uplift in the Greater Caucasus and subsequent propagation of deformation. The
212 south-verging structures due to the southward growth of the frontal Greater Caucasus and the north-
213 verging Adjara-Trialeti FTB structures interfere in the Tbilisi area, creating an outstanding example
214 of incipient collision between two oppositely verging orogenic belts (Alania et al. 2021; Fig. 1).

215

216 **3. Materials and Methods**

217

218 **3.1 Materials**

219

220 Both outcrop samples and cuttings from wells were analysed to assess thermal maturity and TOC
221 content, when available, in the study area (Tables 1 and 2). Samples were analysed using several
222 techniques and results were integrated with published data available for the study area for the first
223 time in this paper (Figs. 1 and 3). Data from Bujakaite et al. (2003), Pupp et al. (2018) and Samsu
224 (2014) are presented in the Results section (Tables 1 and 2), whereas other published data are
225 integrated in the Discussion section.

226 More than two hundred subsurface TOC and Tmax data were derived from seven wells drilled in
227 the eastern Adjara-Trialeti FTB and in the western Kura Basin (where HC exploration is

228 concentrated), from which we extrapolated the Oligocene-Early Miocene interval (Maikop series)
229 in order to have a reference section for interpretation. Pyrolysis results and TOC content estimates
230 were made using Rock-Eval technology and ELTRA Elementar Analyser, respectively. Details on
231 the methods are provided in Pupp et al. (2018) and Samsu (2014). From south to north the wells are
232 (Fig.1):

- 233 I. Kumisi 2
- 234 II. Patardzeuli (SLB)
- 235 III. Satskhenisi 102
- 236 IV. Norio 200
- 237 V. Norio 72
- 238 VI. Ninotsminda 97
- 239 VII. Manavi 12

240 About forty surface samples (Table 1) were analysed or revised in the three geological domains
241 from various stratigraphic intervals: Jurassic (original data and after Bujakaite et al., 2003);
242 Cretaceous (original data); Late Paleocene-Eocene (original data and after Pupp et al., 2018);
243 Oligocene-Early Miocene (original data and after Pupp et al., 2018); Middle-Late Miocene (original
244 data). They were characterised using organic petrography, pyrolysis and micro-Raman analyses on
245 dispersed organic matter, and XRD diffraction on the <2 μm fraction of clay minerals.

246

247 **3.2 Methods**

248

249 *3.2.1 Organic petrography*

250

251 Vitrinite reflectance (VRo%) is generally the best thermal indicator in sedimentary sequences
252 rich in organic matter, and can be correlated with detailed stages of HC generation (Bertrand et al.,
253 2010; Borrego and Cook, 2017; Dow, 1977).

254 Preparation of the samples for optical microscope analysis of dispersed organic matter and
255 vitrinite reflectance required picking of the visible organic matter particles, that can be easily found
256 in the silty and arenaceous fractions, often as vegetable whips (Barnes et al., 1990; Taylor et al.,
257 1998). Once a few grams of rock containing organic matter were selected, they were smoothly
258 crushed in an agate mortar to a medium sand grain size. Obtained powder was placed on a sample
259 holder and incorporated in a two-component epoxy resin. Then specimens were sanded using a
260 Struers LaboPol 5 automatic sanding/polishing machine, with 320, 500 and 1000 grit carborundum
261 sandpaper, and isopropanol lubricant and water as coolers. Then specimens were finally polished
262 with alumina suspensions, with decreasing grain size (1 to 0.3 μm) and microfibre cloths. The
263 routine was completed by polishing with 0.12 μm fumed silica suspension.

264 Vitrinite reflectance analysis was performed at the ALBA (Academic Laboratory of Basin
265 Analysis) of Roma Tre University using a Zeiss Axioskop 40 A microscope equipped with a
266 tungsten halogen lamp (12V, 100W) that produces non-polarized light ($\lambda = 546 \text{ nm}$), Epiplan-
267 Neofluar 50x/1 objective immersed in oil ($n = 1.518$) at a temperature of 23°C, photomultiplier
268 MPS 200 (from J & M Analytik AG), short- and long-wave ultraviolet lamps, coupled with a Canon
269 Power Shot G6 digital camera and a dedicated software for reflectance data acquisition.

270 Before starting measurements, instrument calibration was performed with three reflectance
271 standards. In addition, parasitic light intensity (which varies depending on the intensity of sunlight
272 throughout the day) was measured to allow the software to filter it.

273 The average vitrinite reflectance (VRo%) values were calculated as the arithmetic mean over a
274 minimum of 20 measurements per sample and considered acceptable with a maximum standard
275 deviation of ± 0.06 on the indigenous fragments (Borrego et al., 2006). Each measurement was

276 made on well preserved non-oxidised fragments $>5 \mu\text{m}$ and as far as possible from fractures and
277 pyrite crystals that could decrease or increase true reflectance values, respectively.

278
279 *3.2.2 Micro-Raman spectroscopy on dispersed organic matter*

280
281 Raman spectroscopy is a non-destructive tool to quantitatively evaluate thermal maturity of
282 organic matter from diagenesis to metamorphism (Beysac et al., 2002; Ferralis et al., 2016; Guedes
283 et al., 2010, 2012; Henry et al., 2019; Hinrichs et al., 2014; Lahfid et al., 2010; Liu et al., 2013;
284 Lünsdorf and Lünsdorf, 2016; Mumm and Inan, 2016; Quirico et al., 2005; Schito et al., 2017,
285 2019; Schito and Corrado, 2018; Wilkins et al., 2014; Zhou et al., 2014). Advances in
286 instrumentation and data processing have spurred increased applications, and the technique is now
287 simple, fast and can be performed directly on standard petrographic thin sections or on bulk
288 kerogen.

289 Raman spectra were acquired at the laboratory of experimental volcanology and petrology
290 (EVPLab) of Roma Tre University on standard petrographic thin sections following the procedure
291 described by Beysac et al. (2002) and Lünsdorf et al. (2017). The spectrometer used is a Jobin
292 Yvon micro-Raman LabRam with a backscattered geometry in the range of $700\text{-}2200 \text{ cm}^{-1}$ (1st
293 order Raman spectrum), which uses a grid of 600 meshes per mm and a CCD detector with a
294 maximum magnification of 50x. A Nd-YAG laser with a wavelength of 532 nm (green laser) with a
295 power $< 0.4 \text{ mW}$ was used as energy source. Raman back scattering was then recorded in six
296 repetitions with 20-second steps for each measurement which, together with the use of green lasers
297 and optical filters, helped to reduce the background noise given by the fluorescence of organic
298 matter within acceptable values (Schito et al., 2017). A total of twenty measurements were made for
299 each sample to ensure reproducibility using a $2 \mu\text{m}$ diameter spot at 50x magnification.

300 An automatic approach (Ifors software) was followed for the identification of the number of
301 bands of the Raman spectrum, as illustrated by Lünsdorf et al. (2017) and Lünsdorf and Lünsdorf

302 (2016). The method is based on the STA (Scaled Total Area) parameter, which is a function of the
303 area and maximum intensity of the D and G peaks (Fig. 4). Once this parameter is calculated, using
304 a 532 nm green laser source, it is possible to obtain the maximum burial temperatures by solving
305 the following equation:

$$306 \quad T_{532 \text{ nm}} [\text{°C}] = -8,259 \cdot 10^{-5} \cdot \text{STA}^3 + 3,733 \cdot 10^{-2} \cdot \text{STA}^2 - 6,445 \cdot \text{STA} + 6,946 \cdot 10^2.$$

307

308 *3.2.3 XR diffraction on <2µm grain-size fraction*

309

310 Clay minerals in shales and sandstones undergo diagenetic and very low-grade metamorphic
311 reactions in response to sedimentary and/or tectonic burial. In particular, mixed layers illite-
312 smectite (I-S) and the transformation sequence smectite-randomly ordered mixed layers (R0)-
313 ordered mixed layers (R1 and R3)-illite-muscovite (di-octahedral K-mica) can be used as indicators
314 of the thermal evolution of sedimentary successions (Aldega et al., 2007a, 2007b, 2014; Pollastro,
315 1990).

316 Illite crystallinity (IC) is the measure of the full width at half maximum (FWHM) of the first
317 illite diffraction peak (1 nm = 10 Å) and is a method suitable to detect the anchizonal and its
318 immediate limits, for which it is most accurate (Kübler and Jaboyedoff, 2000).

319 Samples were analysed for qualitative and semi-quantitative analyses of the <2 µm grain size
320 fraction at Roma Tre XRD Laboratory. X-Ray diffraction (XRD) analyses were carried out with a
321 Scintag X1 X-ray system (CuKα radiation) at 40 kV and 45 mA. Randomly oriented whole-rock
322 powders were run in the 2-70° 2θ interval with a step size of 0.05° 2θ and a counting time of 3s per
323 step. Oriented air-dried and ethylene-glycol solvated samples of the <2 µm (equivalent spherical
324 diameter) grain-size fraction were scanned from 1 to 48° 2θ and from 1 to 30° 2θ, respectively, with
325 a step size of 0.05° 2θ and a count time of 4s per step.

326 The illite content in mixed-layers I-S is determined by the delta two-theta method after
327 decomposing the composite peaks between 9–10° 2θ and 16–17° 2θ (Moore and Reynolds, 1997)

328 and by modelling XRD patterns using Pearson VII functions. The R ordering of I-S (Reichweite
329 parameter, R; Jagodzinski, 1949) is determined by the position of the I 001-S 001 reflection
330 between 5 and 8.5° 2 θ .

331 'Illite crystallinity' (IC, also called Kübler Index, KI) measurements are made by first
332 subtracting the background from the raw data, and then applying a profile-fitting method (Lanson,
333 1997). The 10 Å asymmetric illitic multiphase peak was fitted using the Scintag X1 software. Peak
334 shape decomposition was performed on ethylene-glycol preparations using split Pearson VII
335 functions. The peaks identified were rationalized in terms of specific discrete or mixed-layers I-S
336 and/or C-S phases (Lanson, 1997). From fitted data, the crystallinity was determined after
337 calibrating the full width at maximum height (FWHM) of the illite band using Warr and Rice
338 (1994) standards. According to the classification of Kübler (1964), KI values between 0.42 and 0.25
339 correspond to the anchizone (200-300°C) while values lower than 0.25 reflect the onset of epizone
340 at temperatures higher than 300°C.

341

342 3.2.4 Correlation among different indicators of maximum temperature exposure

343

344 Correlation among different thermal indicators and conversion into temperatures are not
345 straightforward since the different factors that drive maturation in sedimentary basins such as
346 thermal regime, sedimentation rate, tectonic subsidence, mineral availability and fluid circulation
347 dissimilarly affect each analytical parameter. Nevertheless, to provide a broad view of the
348 temperature variation among and within the different basins (Fig. 5), we attempted a conversion of
349 organic indicators using the most accepted equations or correlations, i.e. Barker and Pawlewicz
350 (1986) for vitrinite reflectance and vitrinite reflectance equivalent from T_{max} values, and Lünsdorf
351 et al. (2017) for Raman parameters. T_{max} values were preliminary converted into vitrinite
352 reflectance equivalent using Barnard et al. (1981). I% in I-S were converted according to Aldega et

353 al. (2007b) and Merryman and Frey (1999). Throughout the paper, we consider the classical
354 temperature range established for the oil window (ca. 90-120°C, see Hartkopf-Fröder et al. (2015)
355 and references therein). VRo% and I% in I-S were plotted on Hillier et al.'s diagram (1995), where
356 each couple of data can be attributed to different heating rates, that in turn can be ascribed to
357 different geodynamic settings characterised by low (e.g. cratons or foreland basins) to very high
358 (e.g. rift basins) thermal regimes.

359

360 **4. Results**

361

362 Results are analytically provided in Tables 1 and 2 and synthesized in Figs. 3 and 5.

363

364 **4.1 Organic petrography**

365

366 Vitrinite reflectance and main petrographic observations of new surface samples from the
367 Adjara-Trialeti FTB, the Kura Basin and the Kakheti ridge (Table 1) are described below; no data
368 are provided for the Greater Caucasus because the high maturity levels in this domain prevented
369 accurate VRo% determination.

370 The Maastrichtian-Danian to Oligocene-Lower Miocene samples collected in the Kakheti ridge
371 show a decrease in thermal maturity from the oil window (VRo about 0.65%) to the immature stage
372 of hydrocarbon generation (VRo 0.48%). Only three out of five samples provided reliable
373 reflectance values, given the scarcity of OM and the diffuse presence of macerals belonging to the
374 inertinite group, useless for maturity studies. Samples CA20 and CA28 show the most reliable
375 results given the high content of in-situ vitrinite (more than 40 fragments) and the low standard
376 deviation. The presence of yellow and UV-fluorescent sporinites confirms the low maturity level.

377 Three Oligocene-Miocene samples (CA15, CA16 and CA17) from the Kura Basin are
378 characterised by a high content in terrestrial debris with minor amount of inertinite and semifusinite
379 fragments. Vitrinite reflectance measurements indicate as a whole the immature stage of
380 hydrocarbon generation, with VRo values ranging from 0.40 to 0.49% and very low standard
381 deviation values.

382 Finally, seven Paleogene samples from the Adjara-Trialeti FTB were analysed. Organic facies
383 are mainly composed by huminite-vitrinite group fragments with minor amount of inertinite
384 fragments and some sporinites. Pyrite is frequent in globular aggregates of variable sizes. Measured
385 vitrinite reflectance ranges between 0.50 and 0.77% indicating the early and middle oil window
386 stage.

387

388 **4.2 New and revised Tmax and TOC data from pyrolysis**

389

390 Presented Tmax and TOC data generally derive from the Oligocene-Lower Miocene Maikop
391 interval, drilled in seven wells located in the eastern surroundings of Tbilisi (Figs. 1 and 3). They
392 are from unpublished reports produced for the Georgia Oil & Gas Company and from Samsu
393 (2014).

394 Further surface data derive from new sampling along the Kakhети ridge, the Dzirula Massif and
395 the more external units of the Greater Caucasus, whereas data from Pupp et al. (2018) come from
396 the easternmost Adjara-Trialeti FTB and the Kura Basin. They have been selected in order to
397 integrate our new original maturity and obtain the maturity distribution shown in the map of Fig. 3.

398 Tmax values of samples with less than 0.5% TOC were disregarded. Kerogen is mostly type III
399 to type II-III (Fig. 6), with substantial input of terrestrial organic debris. Maturity falls in the
400 diagenetic realm ranging from the immature (<0.5% VRo) to the mid-mature and, rarely, late
401 mature stages of HC generation. Thermal maturity is generally lower in younger (Maikop) with

402 respect to older (Upper Eocene and Cretaceous) stratigraphic intervals (Fig. 6). Analytical data are
403 presented in Tables 1 and 2 and in Figure 7.

404 In detail, the sampled Maikop section in the Kumisi 2 well ranges between 1340 and 1535m
405 depth with a Tmax between 434 and 441°C and a mean value of 439°C, and with TOC ranging
406 between 1.05 and 1.95% with a mean of 1.44% (Fig. 7).

407 In the Patarzeuli (SLB) well, sampled between 360 and 1350m depth, Maikop Tmax ranges
408 between 422 and 440°C with a mean of 431°C, whereas mean TOC is around 1% with minimum
409 and maximum values of 0.64 and 1.25%, respectively (Fig. 7).

410 In the Satskhenisi 102 well, sampled between 200 and 1200m depth, over 80% of the Maikop
411 samples have TOC values higher than 1%, with a minimum value of 0.50% and a maximum one of
412 2.40%. There is no clear TOC trend with depth. Nevertheless, the highest TOC values (>2%) are
413 concentrated towards the bottom half of the section (below 700m of depth) (Fig. 7).

414 The Oligocene interval in the Norio200 well, between 665 and 1233m depth, shows a Tmax
415 between 423 and 429°C with a mean of 427°C, and TOC ranging between 1.75 and 2.19% with a
416 mean of 1.93% (Fig. 7).

417 In the Norio72 well, the Maikop (3625-4510m) yielded Tmax comprised between 418 and
418 435°C with a mean value of 427°C, and TOC ranging between 0.30 and 1.60% with a mean of
419 0.80% (Fig. 7).

420 In the Ninotsminda 97 well the Lower Oligocene interval of the Maikop (2330-2360m) yielded
421 Tmax values between 421 and 424°C with a mean of 422°C. TOC ranges between 0.63 and 0.82%,
422 with a mean of 0.76%.

423 The Manavi 12 well crosses the Maikop between 3800 and 3920m depth with Tmax indicating
424 the immature stage of HC generation (between 407 and 431°C with a mean of 424°C), and TOC
425 between 3.30 and 5.40% with a mean of 4.20%.

426 New surface samples collected along the Kakheta ridge and the frontal tectonic units of the
427 Greater Caucasus show higher Tmax values in the latter ones (mainly >450°C in Upper Eocene

428 rocks and 445°C in the Cenomanian sample CA35). In the Kakheti ridge Late Eocene Tmax ranges
429 between 434 and 442°C and Oligocene-Miocene Tmax ranges between 421 and 427°C, suggesting
430 a lateral trend of decreasing maturity moving from the WNW to ESE and an increase of maturity
431 from younger to older strata.

432 Original surface data from the Cretaceous unconformably lying on top of the Dzirula massif
433 indicate a Tmax lower than 437°C.

434 Surface sections studied by Pupp et al. (2018) between the Adjara-Trialeti FTB and the Kura
435 Basin provide reliable Tmax data for the Late Eocene-Oligocene (446-448°C), Late Oligocene-
436 Miocene (429-435°C) and Miocene (<437°C) intervals, showing an overall maturity decrease from
437 older to younger units. Oligocene sample CA4, collected in the Akhaltsikhe depression, yielded a
438 Tmax value of 430°C.

439

440 **4.3 Raman spectroscopy on dispersed organic matter**

441

442 Measurements of Raman spectra on dispersed organic matter were focused on samples from the
443 Greater Caucasus, where metamorphic temperatures make vitrinite reflectance data less reliable for
444 thermal maturity assessments. Here, Raman spectra show a clear-cut temperature increase going
445 from the younger to the older chronostratigraphic units as well as toward the axis of the orogen
446 (Table 1). This trend is illustrated in Fig. 4, where spectra from Lower Jurassic to Middle
447 Cretaceous samples collected along a N-S transect running parallel to the Georgian Military Road
448 are shown. Upper Jurassic sample CA39 crops out a few tens of km to the ESE of it and thus its
449 thermal evolution should be considered separately from the other samples; for this reason, the
450 Raman spectrum of this sample is not included in Fig. 4.

451 Spectra from Aptian to Cenomanian successions (samples CA36 and CA37), as well as in Upper
452 Jurassic sample CA39, are characterised by a broad asymmetric D band at 1350cm^{-1} with lower

453 intensities with respect to the G band at 1600cm^{-1} . Such spectra are typical of very low
454 metamorphic conditions, as outlined by the temperature range from the Ifors software ($219\text{-}235^\circ\text{C}$).
455 Moving toward the Berriasian-Hauterivian sample (CA38) the D band shows relatively higher
456 intensities and the bands that underlines the “saddle” between the D and G bands tend to disappear.
457 These changes reflect a temperature increase up to $292 \pm 7^\circ\text{C}$, near the boundary between
458 anchizone and epizone ($\sim 300^\circ\text{C}$ according to Kübler, 1964). Finally, in the Jurassic sample
459 (Toarcian-Aalenian, CA40) the intensity of the D band is higher than that of the G band, which
460 shows a marked asymmetry at 1620cm^{-1} due the presence of the D2 band. All these features
461 correspond to an average temperature of $379 \pm 9^\circ\text{C}$.

462

463 **4.4 XRD on $<2\ \mu\text{m}$ grain-size fraction**

464

465 Data obtained from clay mineralogy analyses are shown in Table 1. The interpretation of
466 diffractograms on air-dried and glycolated samples are expressed as abundance in percentage to
467 provide the composition of the $<2\ \mu\text{m}$ fraction. Moreover, where illite-smectite (I-S) mixed layers
468 are present, the percentage of illite in mixed layers (I% in I-S), which can be interpreted in terms of
469 maximum temperature, and the parameter R (Reichweite index) are provided. IC (illite crystallinity)
470 is provided for samples where only illite (without I-S mixed layers) is present.

471 In the Greater Caucasus, three samples (CA36, CA37 and CA38) contain 8 to 28% of I-S mixed
472 layers, together with illite (1-88%) and chlorite (6-71%). These samples have a R3 stacking order of
473 illite-smectite mixed layers and I% in I-S ranging from 80% (Cenomanian, CA36) to 88% (Aptian-
474 Albian, CA37), indicating the late diagenetic realm. The remaining two samples from the Greater
475 Caucasus, CA40 (Toarcian-Aalenian) and CA39 (Oxfordian-Tithonian), contain illite, chlorite and
476 chlorite-smectite mixed layers. According to Kübler's (1964) classification, IC values of the Lower
477 Cretaceous (CA38) and the Lower-Middle Jurassic (CA40) samples indicate respectively anchizone

478 and epizone conditions whereas the Upper Jurassic sample (CA39), located >70 km to the ESE,
479 reached only late diagenetic conditions.

480 The six samples from the Kakheti ridge of the Kura Basin contain I-S mixed layers (generally
481 >20%), illite (30-65%), chlorite (2-30%) and kaolinite (10-30%), with the exception of the
482 Maastrichtian sample where these last two phases are absent. Kaolinite -indicative of intense
483 weathering- is virtually absent in samples from other domains. I% in I-S mixed layers ranges
484 mainly between 70 and 75%, with R1 stacking order. Two samples (CA28, Maastrichtian, and
485 CA19, Oligocene-Early Miocene), collected near two fault zones, show much lower I% in I-S
486 mixed layers (18 and 15%, respectively) probably due to alteration that could increase the smectite
487 percentage.

488 Two Miocene samples from other areas of the Kura Basin provided reliable XRD results,
489 indicating low diagenetic conditions. CA14 (Late Miocene) contains mainly smectite (85%) and
490 minor amounts of illite and chlorite (10 and 5%, respectively). CA16 (Early Miocene) contains
491 illite, chlorite, illite-smectite mixed layers (<20%) with a percentage of illite in mixed layers of
492 45% and R0 stacking order. The abundance of kaolinite (>30%) indicates weathering processes.

493 Almost all of the eight samples from the Paleogene section of the Adjara-Trialeti FTB contain
494 chlorite-smectite mixed layers, nearly absent in the other domains, with percentages ranging from
495 about 20 to 80%. Illite is ubiquitous with highly variable percentages (5-72%), and chlorite is
496 frequently present, with a lower percentage in the Paleocene-Early Eocene sample (CA13, 6%) than
497 in the Eocene section (28-33%). I-S mixed layers are present only in four samples in the Eocene
498 section, with percentages between 24 and 40%. I% in I-S mixed layers is highly variable, with two
499 samples (CA11 and CA7, Early and Late Eocene, respectively) yielding a percentage between 70
500 and 76% and R1 stacking order, and two samples (CA10 and CA9, Middle and Late Eocene,
501 respectively) with I% between 27 and 32% and R0 stacking order.

502

4.5 *Extrapolation of paleotemperature ranges and integration with pre-existing datasets*

505

506 Organic indicators (VRo% and Tmax from pyrolysis) from virtually all of the Paleocene-
507 Miocene samples of the Adjara Trialeti FTB point consistently to the diagenetic realm (early to
508 mid-mature stages of hydrocarbon generation), thus never exceeding about 115°C. I% in I-S and R
509 ordering show, on the other hand, contrasting values (Table 1). Considering organic results, we
510 suggest that the two samples (CA8 and CA9) with low I% in I-S mixed layers and R0 ordering have
511 probably been affected by alteration processes (hence these two samples were not plotted on Fig. 8).
512 In the Hillier's (1995) diagram (Fig. 8) the correlation between VRo% and I% in I-S for sample
513 CA11 indicates a medium heating rate during back-arc extension in the eastern Adjara-Trialeti
514 basin, before inversion.

515 The three Oligocene-Miocene samples from the Kura Basin yielded coherent paleotemperatures
516 between 60 and 80°C, based on VRo% and I% and stacking order of I-S mixed layers (Fig. 5). The
517 correlation between VRo% and I% in I-S for sample CA16 indicates a moderate heating rate typical
518 of relatively cold basins (Fig. 8). This is further confirmed by Tmax values from the Oligocene-
519 Early Miocene Maikop unit, or part of it, in the wells near Tbilisi (Figs. 1, 3 and 6, Table 2),
520 indicating that the maturity trend with depth defines cold conditions, and by Pupp et al. (2018) who
521 define a maturity gradient of 0.08%/km (Δ VRo%) in the Maikop series, typical of cold foreland
522 basins. The new pyrolysis results (Tmax, TOC and HI; Tables 1 and 2, Figs. 6 and 7) also show that
523 the organic matter is mainly of type III to type II-III, thus rich in terrestrial input: this evidence and
524 the maturity level reconstructed are in agreement with published results from wells and outcrop
525 samples in the western Kura Basin (Pupp et al., 2018; Samsu, 2014).

526 The thermal maturity of the Cretaceous sample (CA18) collected at the surface from the
527 unconformable cover of the Dzirula Massif indicates shallow burial.

528 As for the Kakheta ridge, all reliable data fall in the diagenetic realm showing a general increase
529 of paleotemperatures derived from organic indicators (T_{max} and V_{Ro}%) from 40-50°C to about
530 100°C moving from younger (Oligocene-Lower Miocene) to older (Lower Cretaceous)
531 lithostratigraphic units (Figs. 5 and 6). XRD analysis of the <2 μm fraction indicates the presence of
532 I-S mixed layers with an illite content of more than 70% and R1 ordering. These results suggest a
533 much higher thermal maturity with respect to V_{Ro}% and T_{max} that can be interpreted as detrital
534 clay contamination. On the other hand, two samples collected close to thrusts show a much lower
535 illite content in mixed layers (<20%) probably due to post-diagenetic smectite enrichment caused
536 by fluid circulation. Hillier's diagram (Fig. 8) indicates a low to moderate heating rate, typical of
537 relatively cold basins (samples CA27-CA20).

538 In the Greater Caucasus, Raman spectroscopy and XRD (KI and I% in I-S) in the axial part of
539 the belt, and T_{max} in the southern foothills provided new reliable results (Figs. 5 and 9). In the
540 Lower Jurassic section sampled on the highest thrust sheet of the Greater Caucasus close to the
541 Russia-Georgia border, the maximum paleotemperatures derived from XRD and Raman are
542 comprised between 300 and 400°C, with a more refined range given by Raman spectroscopy (360-
543 400°C, Figs. 4 and 5). Raman results should be privileged as KI-derived paleotemperatures result
544 from a discontinuous correlation scale (Hoffman and Hower, 1979) whereas the equation used to
545 calculate temperatures from Raman parameters (Lünsdorf et al., 2017) provides a more detailed
546 resolution. Additional KI data (Bujakaite et al. (2003) confirm that maximum paleo-temperatures
547 were higher than 300°C. These results are further corroborated by thermochronological results from
548 Vasey et al. (2020), indicating that the basement in the axial zone of the Greater Caucasus near Mt.
549 Kazbek (Fig. 1) underwent very fast cooling since about 10-8 Ma from temperatures higher than
550 ~250°C.

551 Moving to the footwall of the highest thrust, KI data indicate a decrease in paleotemperatures
552 from more than 300 to about 250°C going upsection from north (Lower Jurassic) to south (Upper
553 Jurassic) (Bujakaite et al., 2003).

554 New Raman, KI, Tmax and I% in I-S data from the bottom of the Cretaceous succession indicate
555 a well constrained range between 280 and 300°C, that decreases to 180-260°C moving towards the
556 uppermost Lower Cretaceous and Cenomanian units. Here the percentage of illite probably slightly
557 underestimates paleotemperatures for lithological reasons (abundant carbonate cement) whereas R3
558 stacking order suggests paleotemperatures partially superposing with Raman ones.

559 In the southern foothills of the Greater Caucasus, an area dominated by thin-skinned thrust
560 sheets, Tmax values suggest a paleotemperature range between 100 and 160°C in the Upper Eocene
561 rocks, corroborated by the Tmax vs HI relationships (Fig. 6) which indicate a late mature to
562 overmature stage of HC generation. These results show that the youngest stratigraphic units in the
563 orogenic belt are characterised by the lowest maturity level.

564 Overall, the integration of the new data presented here with pre-existing datasets describes a
565 progressive increase of thermal maturity from the southernmost foothills to the highest peaks of
566 Greater Caucasus in Russian territory, i.e. from younger to older strata (Figs. 5 and 9).

567

568 **5. Discussion**

569

570 **5.2 Thermal histories**

571

572 The thermal maturity distribution for the three geological domains characterising the study area
573 is represented in Fig. 5 and along the profile in Fig. 9. In the Paleocene-Lower Eocene section of
574 the Adjara-Trialeti FTB, thermal maturity decreases from south (Tetri Tsqaro) to north (Mtskheta),
575 from the peak of the oil window to the early mature zone. The difference in thermal maturity is
576 relatively small but could be explained with the possible effect of tectonic thickening due to the
577 inverted E-W fault running in the area of Amlevi, evolving into a low-angle thrust in its present-day
578 eroded portion (Figs. 1, 3 and 9).

579 The thermal maturity of the Middle-Upper Eocene section does not show significant relevant
580 lateral variations. It ranges from the immature stage of HC generation to the oil window and is
581 slightly lower in comparison with the Paleocene-Eocene section.

582 The thick Oligocene-Lower Miocene (Maikop) section of the Adjara-Trialeti FTB reached the
583 early mature oil window, whereas it is immature in the central portion of the orogen [samples CA1
584 and CA2 (Pupp et al., 2018)]. Further to the north, where the Maikop series is overlain by Middle-
585 Upper Miocene sediments in the frontal synform of the Adjara-Trialeti FTB (Figs. 3 and 9), thermal
586 maturity increases again to the early mature stage. The thermal immaturity of the Maikop section in
587 the Kumisi 2 well could be explained with the closure of the former Adjara-Trialeti back-arc basin
588 towards the SE, hence to the lower amount of experienced burial (Fig. 3).

589 The positive correlation between thermal maturity and stratigraphic age suggests the limited
590 effect of tectonic loading, apart for the southern part of the cross-section in Fig. 9. Maximum
591 temperatures recorded in the Adjara-Trialeti domain are consistent with the thickness of the rift
592 basin fill and hence resulted from sedimentary burial. The thermal maturity trend reconstructed in
593 the Adjara-Trialeti FTB thus provides independent evidence that the dominant deformation style
594 within the orogen is positive inversion of the former extensional faults, as already pointed out in
595 previous works (Alania et al., 2020; Banks et al., 1997; Gusmeo et al., 2021; Sosson et al., 2016),
596 rather than low-angle thrusting, a process which would have resulted in a higher thermal maturity
597 due to tectonic overburden.

598 Based on our results, suggesting maximum paleotemperatures between about 70 and 120°C, and
599 assuming an average geothermal gradient of 25-30°C/km, the total eroded basin fill of the eastern
600 Adjara-Trialeti Basin broadly ranges between 2 and 4.5 km. These rough estimates are in agreement
601 with Pupp et al. (2018), who estimated ~3/3.5 km of eroded Miocene section in the eastern Adjara-
602 Trialeti FTB.

603 Maturity data from the subsurface of the Kura Basin are available only for the Maikop series and
604 indicate an immature to early mature stage of HC generation (Fig. 6) acquired during progressive

605 burial by the Middle-Late Miocene stratigraphic section (Pupp et al., 2018). Surface samples within
606 the Kura Basin are also characterised by a low maturity level (Fig. 5), with maximum
607 paleotemperatures in the 60-80°C range. Assuming 25-30°C/km as average geothermal gradient no
608 more than 2-3 km of basin fill could have been eroded in the deformed Kura Basin. The low
609 maturity of both surface and subsurface samples indicates a cold thermal regime, typical of flexural
610 foreland basins (Fig. 8).

611 In the Kakheti ridge thermal maturity changes from the early mature to the immature stage of
612 HC generation from the Lower Cretaceous to the Maikop sections, respectively, suggesting that
613 maturity was acquired, as in the Kura Basin, as a consequence of sedimentary burial in a quite cold
614 regime, with limited overthrusting effects on the organic matter indicators, whereas thrust-related
615 fluid circulation may have affected clay mineralogy results (see Section 4.5). The Kakheti ridge was
616 exhumed earlier than the rest of the Kura Basin (Fig. 10) and maturity is generally lower than in the
617 wells drilled to the south of the ridge in the eastern and southern surroundings of Tbilisi, because of
618 the lower amount of burial experienced. Our results indicate maximum paleotemperatures between
619 40 and 110°C, which roughly translate into 1-3.5 km of eroded basin fill (assuming 25-30°C/km of
620 geothermal gradient).

621 The comparison of the thermal maturity distribution in the Adjara-Trialeti FTB (in its Paleocene-
622 Miocene section), the Kura Basin (in its Oligocene-Miocene section) and the Kakheti ridge (in its
623 Upper Cretaceous-Oligocene section), indicates that:

624 • Thermal maturity trends in the Adjara-Trialeti FTB and in the Kura Basin/Kakheti ridge are
625 similar, showing an increase in maturity, from younger to older units, from immature to mid-mature
626 oil window. The thickness of the Paleocene-Oligocene sections in the two domains is however
627 different (i.e. 2-3 km thicker in the Adjara-Trialeti FTB), supporting the hypothesis that the final
628 cumulative thermal maturity cannot be due to the same burial/thermal evolution through time. As
629 recognized by Pupp et al. (2018), the maturation of the Oligocene-Lower Miocene source rock in

630 the Kura Basin in the Tbilisi area is due to sedimentation and tectonic thickening of the Neogene
631 basin fill (Nemčok et al., 2013), thus acquired later than in the Adjara-Trialeti inverted basin.

632 • Peak temperatures in the Adjara-Trialeti basin must have been reached during back-arc
633 basin evolution, probably due to enhanced sediment accumulation driven by subsidence in
634 Paleocene-Early Miocene times when the Kura Basin substratum represented a relative structural
635 high. This is also an indirect proof that in Middle-Late Miocene times the Adjara-Trialeti domain
636 was no longer undergoing subsidence and was being exhumed. At the same time the adjacent Kura
637 Basin was experiencing strong subsidence and sedimentation. This is supported by independent
638 thermochronological data and models (Gusmeo et al., 2021) which define the onset of exhumation
639 in the Adjara-Trialeti FTB at around 14-12 Ma.

640 • Compared with the Adjara-Trialeti FTB and the rest of the Kura Basin, the Kakheti ridge
641 shows a slightly lower maturity in time-equivalent stratigraphic units (immature stage in the
642 Oligocene section), confirming that pre-Middle Miocene burial did not cause significant maturation
643 of the Maikop section. Furthermore, in Middle-Late Miocene times the fold-and-thrust belt
644 shortened and started exhuming, with scarce accumulation of syn-tectonic sediments, confined in
645 shallow thrust-top basins, later on sutured by Plio-Pleistocene sediments (Fig. 10a; Alania et al.,
646 2017).

647 The highest paleotemperatures were recorded in the Greater Caucasus, with thermal maturity
648 values spanning from the upper oil/gas generation window up to 200-250°C in the youngest units
649 (the Upper Cretaceous-Eocene section at the front of the belt), to anchizone and epizone in the
650 Jurassic section in the inner portion of the chain, where recorded maximum paleotemperatures
651 exceed 300°C. Presently there are no constraints on the timing of acquisition of such a thermal
652 signature; further work has still to be done in order to solve this issue.

653 These paleotemperatures, if we tentatively assume an average geothermal gradient of 25/30°C,
654 suggest that a minimum of 3.5 km of basin fill has been removed in the southern foothills of the

655 Greater Caucasus, and up to 12.5 km of section may have been removed in the axial zone. These
656 rough estimates are broadly in agreement with reconstructions of the total basin fill of the former
657 Greater Caucasus Basin, although a contribution due to tectonic nappe emplacement to thermal
658 maturity cannot be excluded (Adamia et al., 2011b; Saintot et al., 2006; Vincent et al., 2016).

659 Time-equivalent lithostratigraphic units in the Kakheti ridge and the Greater Caucasus (e.g.,
660 Oligocene-Late Eocene) underwent a different thermal evolution, suggesting that the Kakheti ridge,
661 though appearing laterally contiguous with the southern Greater Caucasus orogenic edifice (Figs. 1
662 and 11), is not really akin to it. Thermal maturity trends in this fold-and-thrust belt are much more
663 similar to the ones in the Kura Basin, thus confirming that the Kakheti ridge represents a highly
664 tectonised and more deeply exhumed portion of that basin.

665

666 **5.2 Regional geological evolution**

667

668 In conclusion, the geological evolution of the sector comprised between the Adjara-Trialeti FTB
669 and the Greater Caucasus can be described as follows (Fig. 11).

670 In the Late Eocene, both Adjara-Trialeti and Greater Caucasus basins were experiencing
671 extensional subsidence, but with different degrees of extension and thickness of sedimentary fill.
672 The Greater Caucasus Basin was characterized by a thin and intruded continental crust in its axial
673 zone, and by a sedimentary succession at least 7-9 km thick (Adamia et al., 2011a, 2011b; Ershov et
674 al., 2003; Saintot et al., 2006). The Adjara-Trialeti Basin was underlain by a thicker and less
675 intruded continental crust and filled by a thinner sedimentary succession (with thickness variable
676 along-strike) (Adamia et al., 2011b, 2017; Gamkrelidze et al., 2019; Nemčok et al., 2013;
677 Okrostsvaridze et al., 2018; Yılmaz et al., 2000, 2014). Between the two basins, a structural high
678 was located, characterized by a very thin sedimentary succession overlying the basement: this

679 structural high will later crop out as the Dzirula Massif, and more to the east it will represent the
680 basement upon which the Kura Basin and the Kakhetai ridge will develop.

681 The present-day situation witnesses both former basins closed and shortened through positive
682 inversion, with the Kura Basin and the Kakhetai ridge trapped between them. Growth of the Adjara-
683 Trialeti FTB started in the late Middle Miocene (Gusmeo et al., 2021), while timing of Greater
684 Caucasus growth is still debated (see section 2; Avdeev and Niemi, 2011; Forte et al., 2014; Vasey
685 et al., 2020; Vincent et al., 2011, 2020). These two orogens are underlain by a thick continental
686 crust, 35 to 45 km below the Adjara-Trialeti belt and 50-55 km in the Greater Caucasus (Adamia et
687 al., 2017; Brunet et al., 2003; Ershov et al., 2003; Motavalli-Anbaran et al., 2016). The Kura Basin
688 was flexured during convergence, mainly in Miocene times, when a thick pile of sedimentary rocks
689 eroded from both the adjacent orogenic belts was deposited within the basin (Adamia et al., 2010;
690 Pupp et al., 2018), causing maturation of the Maikop series, as evidenced in this study. The Maikop
691 succession in the Kakhetai ridge is characterised by a slightly lower maturity degree because the
692 ridge was experiencing shortening and exhumation while the Kura Basin was still experiencing
693 flexural subsidence and sedimentary burial.

694 Our thermal maturity results, integrated with published ones and with the structural framework
695 of the study area, can be interpreted in the broader context of the Arabia-Eurasia collision zone. The
696 tectonic evolution reconstructed in this paper, schematically summarised in Fig. 11 and derived
697 from our thermal maturity data and independent stratigraphic and structural constraints, describes a
698 net change in the dominant stress field which occurred in Middle Miocene times, when extensional
699 tectonics was replaced by compression and the Adjara-Trialeti basin was structurally inverted
700 (Alania et al., 2017; Forte et al., 2014; Gusmeo et al., 2021; Sukhishvili et al., 2020; Tari et al.,
701 2018). Miocene shortening occurred also in the Greater Caucasus, although the timing of its
702 inception is a matter of debate (Cowgill et al., 2016; Vasey et al., 2020; Vincent et al., 2020). This
703 compressional geodynamic regime continues to the present day (Reilinger et al., 2006; Sokhadze et
704 al., 2018; Tibaldi et al., 2019).

705 Inception of compressional tectonics in the study area is coeval with the Arabia-Eurasia hard
706 collision along the Bitlis-Zagros suture zone (Cavazza et al., 2018; Okay et al., 2010). At the same
707 time, wide areas of the suture-zone hinterland, comprising segments of the eastern Pontides, the
708 Caucasus *s.l.*, the Talysh belt, and the Alborz range, were also subjected to deformation (Albino et
709 al., 2014; Axen et al., 2001; Ballato et al., 2011, 2016; Cavazza et al., 2017, 2019; Gavillot et al.,
710 2010; Gusmeo et al., 2021; Koshnaw et al., 2017, 2020; Madanipour et al., 2017; Su and Zhou,
711 2020; Tibaldi et al., 2017). From this viewpoint, the results shown in this paper further support the
712 hypothesis that the compressional stresses associated to the Arabia-Eurasia hard collision might have
713 been transmitted to the north over long distances, causing far-field deformation in a wide area of the
714 hinterland.

715 A fundamental unresolved issue is the exact timing of growth and structural evolution of the
716 Greater Caucasus, which according to our results has recorded a range of paleotemperatures much
717 higher than those recorded in the other domains. The high maturity level in the axial zone of the belt
718 and in the Jurassic-Cretaceous stratigraphic successions, characterised by a thick-skinned tectonic
719 style, can be ascribed with confidence mostly to sedimentary burial. Middle Jurassic magmatic
720 activity (Adamia et al., 2011a; Saintot et al., 2006) may have in part contributed to the very high
721 paleotemperatures recorded in the oldest, most mature samples. Both the maturity trend and the
722 structural style of deformation are in agreement with a positive inversion of the former extensional
723 faults (similarly to the Adjara-Trialeti FTB). The fairly high paleotemperatures (100-160°C)
724 experienced by the Upper Eocene succession of the Greater Caucasus southern foothills require a
725 different explanation. Such maturity level can result from either (i) sedimentary burial underneath a
726 thick succession of younger sedimentary rocks, later almost totally eroded, or (ii) thrust-related
727 tectonic burial. The discrimination of the dominant contribution (sedimentary, tectonic, or both) to
728 the thermal maturity of this part of the belt has crucial implications for a precise reconstruction of
729 the Cenozoic development of the Greater Caucasus.

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6. Conclusions

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This paper provides new constraints on the thermal evolution and the structural styles of a wide area in eastern Georgia, where three geologic domains can be identified in the hinterland of the Arabia-Eurasia collision zone. Two domains derived from positive inversion of former rift basins (i.e. Adjara-Trialeti FTB and Greater Caucasus), the third is comprised between them and characterised by thin-skinned deformation (i.e. Kura Basin and Kakheti ridge). Integrating newly acquired and published thermal maturity indicators we were able to (i) define the maximum temperatures experienced by the sedimentary successions and (ii) to elucidate the tectonic evolution of the area of study during convergence, including the role played by inherited pre-shortening structures.

The results indicate that the Cretaceous-to-Lower Miocene sedimentary units in the Adjara-Trialeti FTB and in the Kura Basin have a similar thermal maturity degree, comprised in the oil window, whereas time-equivalent successions in the Kakheti ridge are slightly less mature (immature to early mature) and the Middle-Late Miocene section of the Kura Basin is immature. The similar thermal maturity in the same stratigraphic units was acquired during back-arc basin evolution in the Paleogene in the Adjara-Trialeti FTB, and during flexure and sedimentary burial, associated to convergence, in the Miocene in the Kura Basin/Kakheti ridge. The Greater Caucasus is characterised by a much higher maturity level, increasing from the southern foothills to the axial zone, where it reaches the low metamorphic realm. Such a maturity probably represents the cumulative effect of both sedimentary burial during extensional evolution and tectonic overburden during compressional deformation.

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Figures

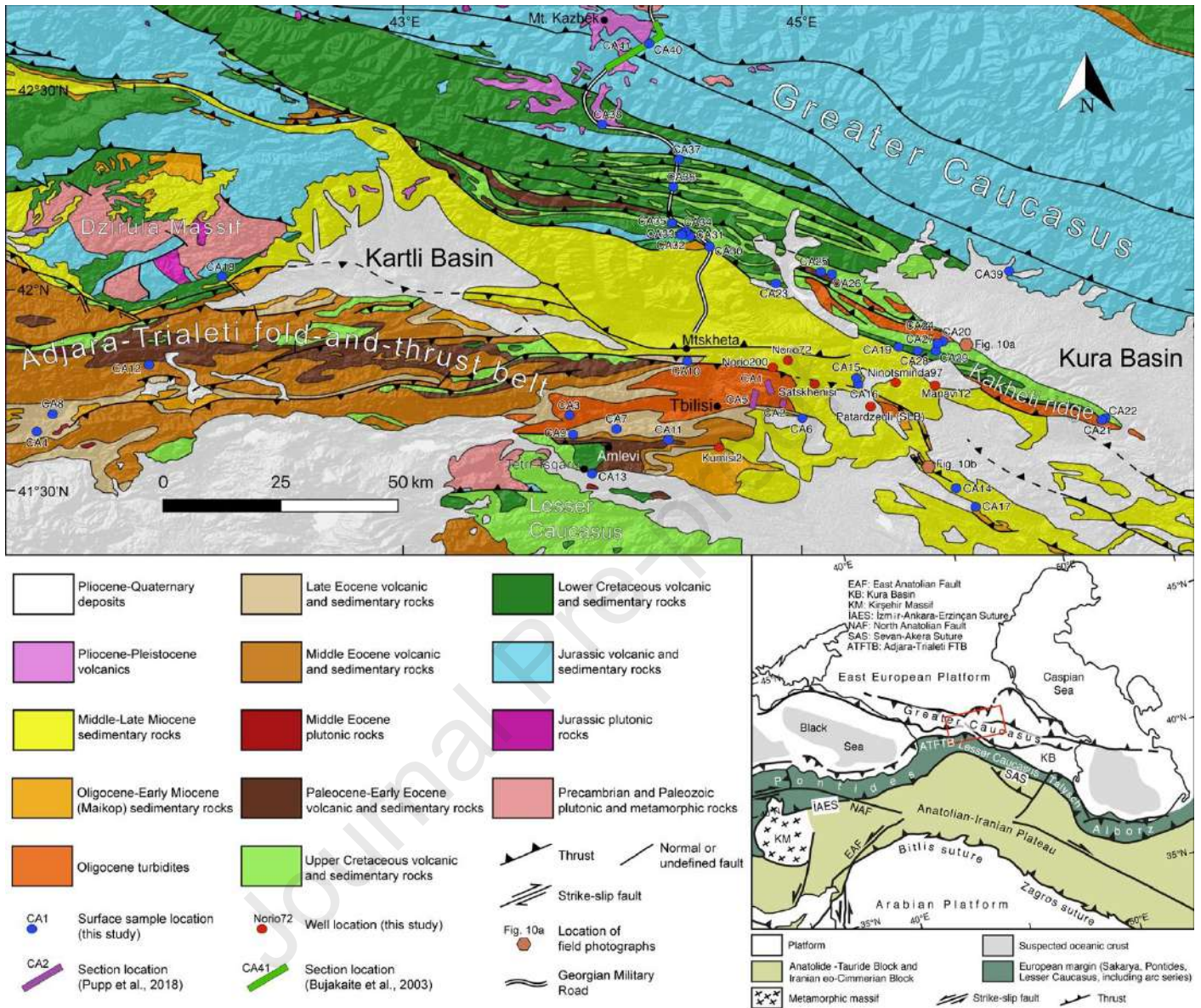


Figure 1: Geological map of the study area modified after Adamia (2004) and Gusmeo et al. (2021) with discussed sample sites and sections. Sample numbers refer to Tables 1 and 2. Locations of field photographs in Figure 10 are also shown. Lower-right inset: geodynamic setting of the collision zone between Eurasia and Arabia (after Cavazza et al., 2019; Sosson et al., 2010). Red rectangle indicates position of Figure 3.

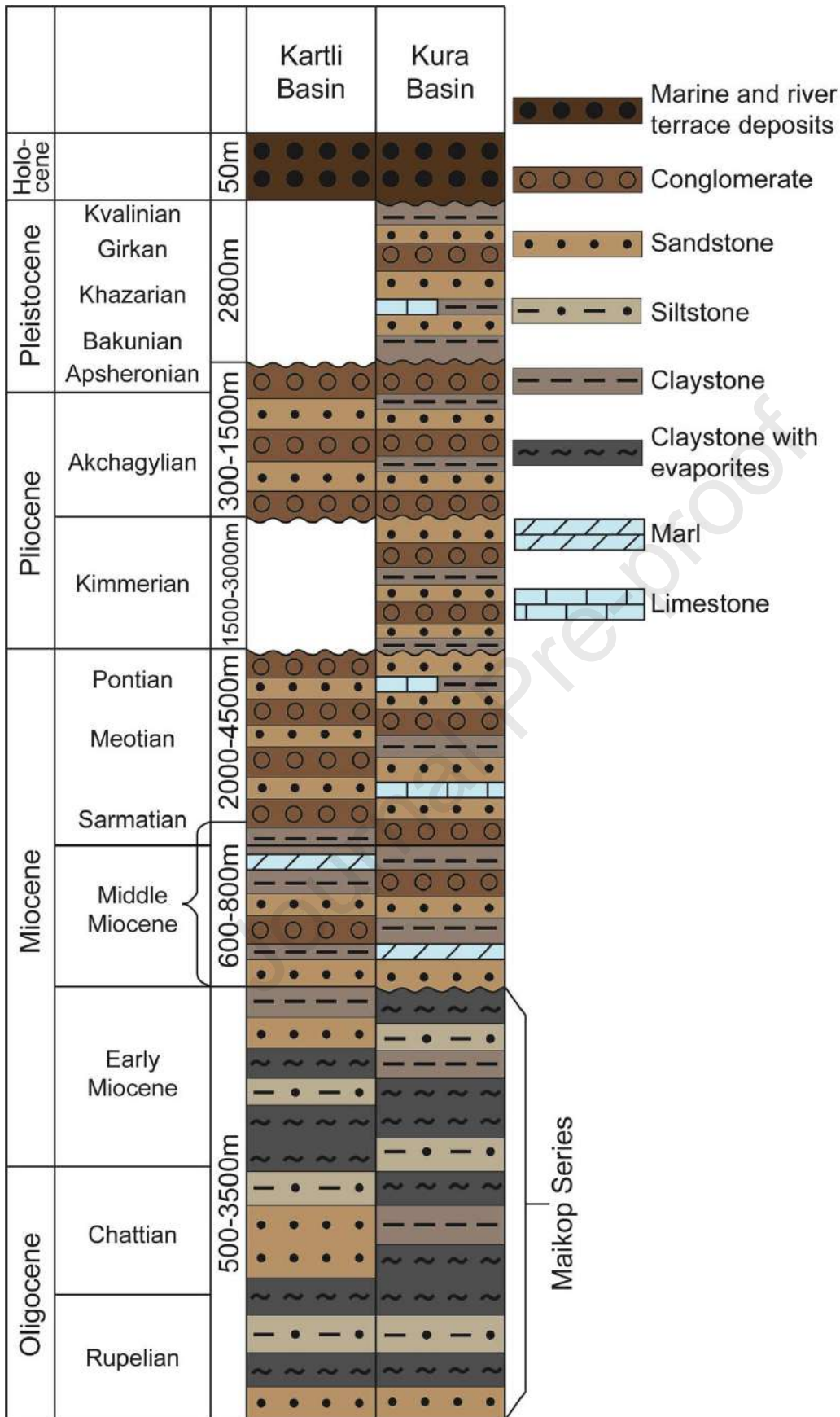


Figure 2: Schematic chrono-lithostratigraphic columns of the Kartli and Kura basins, after Adamia et al. (2010, 2011b) and Pupp et al. (2018).

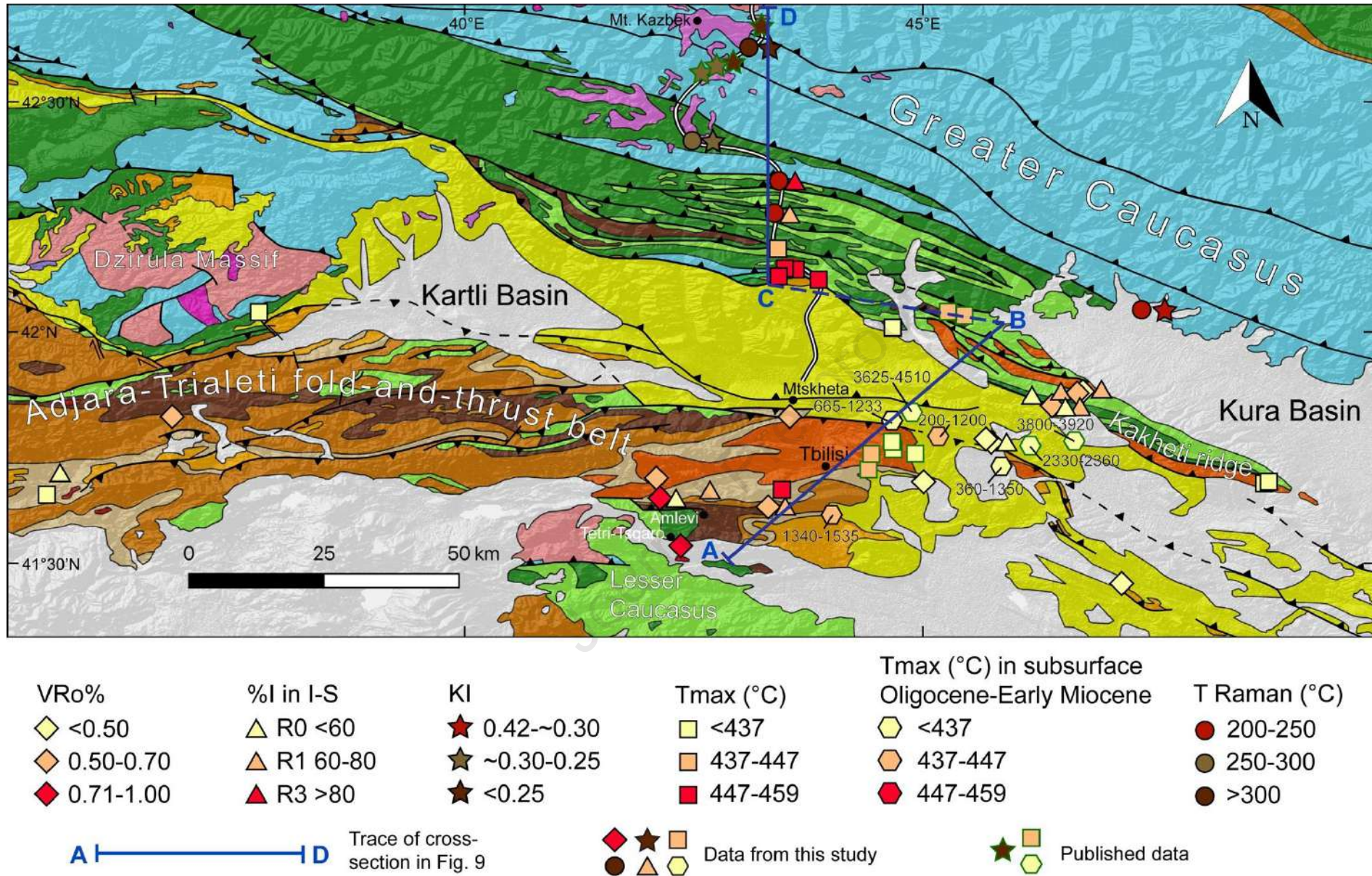


Figure 3: Geological map with original and published synthetic paleo-thermal maturity datasets. Depths of Tmax data from deep wells are shown. Base map modified after Adamia (2004) and Gusmeo et al. (2021). Colours and symbols of base map as in Figure 1.

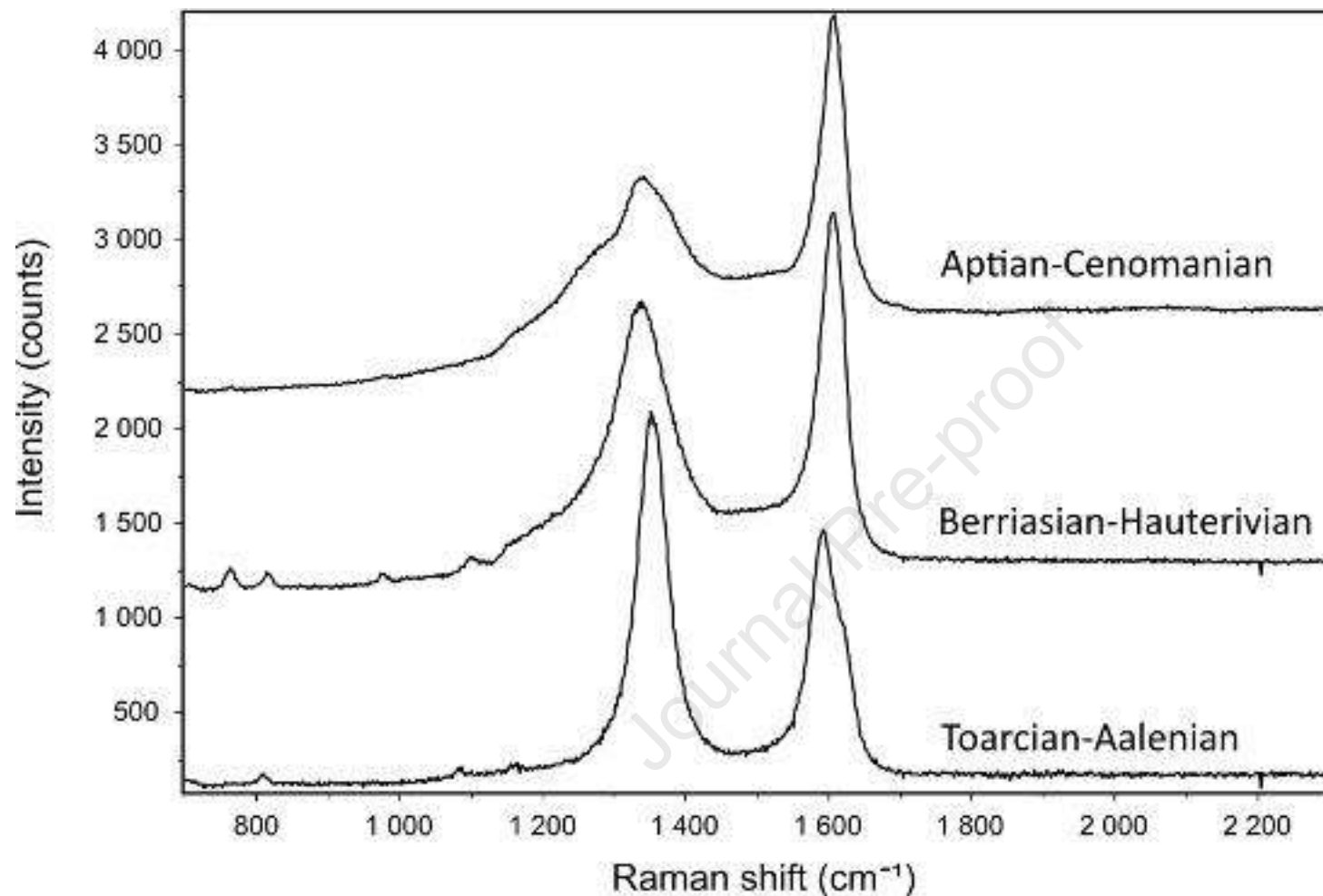


Figure 4: Selected Raman spectroscopy spectra on dispersed organic matter for samples collected across the Georgian Military Road in the Greater Caucasus. The upper spectrum refers to samples CA36 and CA37 (Aptian-Cenomanian), which are very similar, the central spectrum refers to sample CA38 (Berriasian-Hauterivian), and the lower spectrum refers to sample CA40 (Toarcian-Aalenian). D peak is around 1350 cm⁻¹ and G peak is around 1600 cm⁻¹. See text for details.

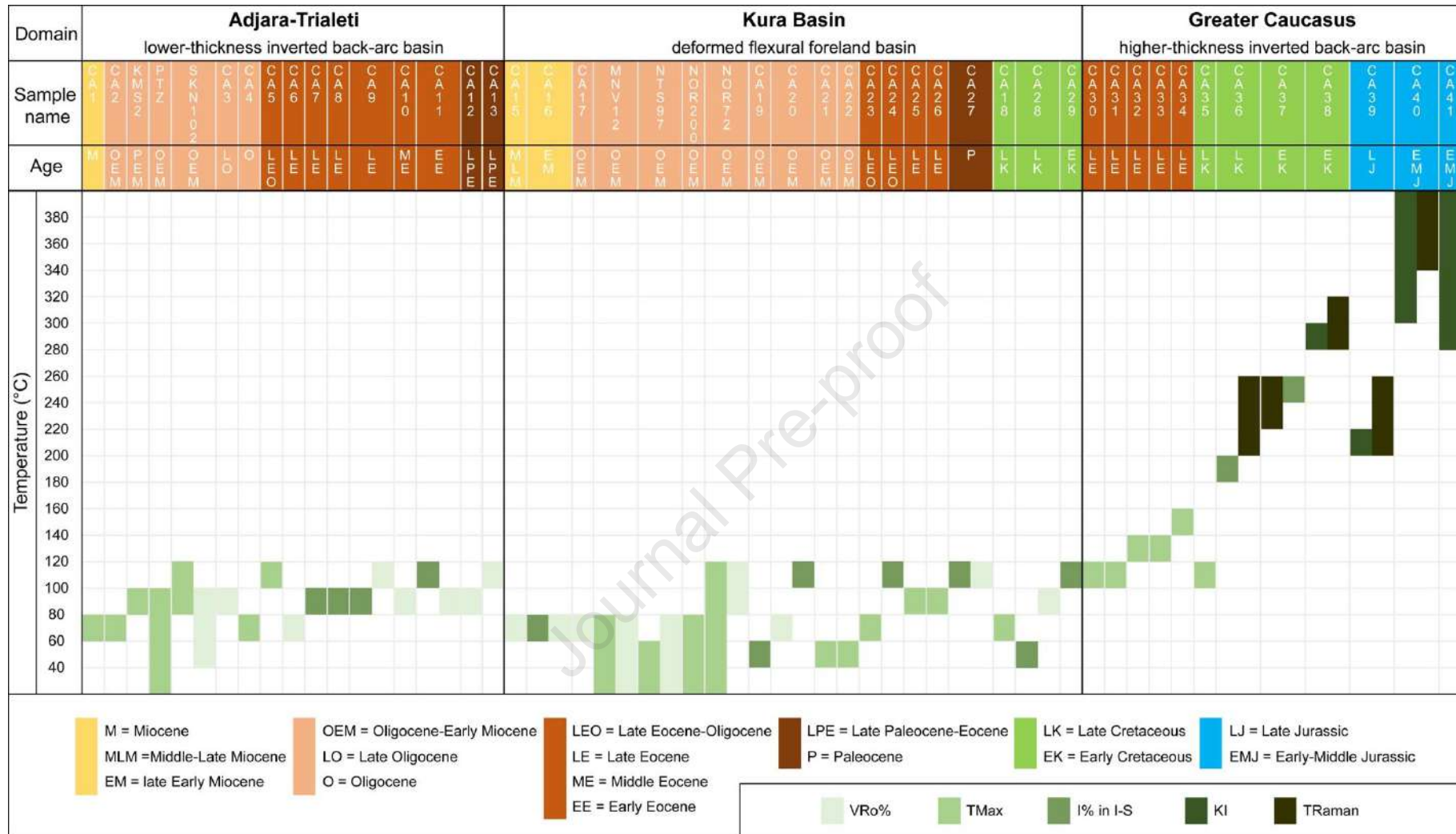


Figure 5: Correlation scheme of paleotemperatures derived from original and published samples according to VRo%, illite% and R number in illite-smectite mixed layers and Tmax with TOC >0.5%. Paleotemperatures from VRo% are derived after Barker and Pawlewicz (1986) equation; from I-S after Hoffman and Hower (1979) and from Tmax after Barnard et al., (1981). In each domain, samples are listed -from left to right- first in chronological order then in geographical (south to north) order.

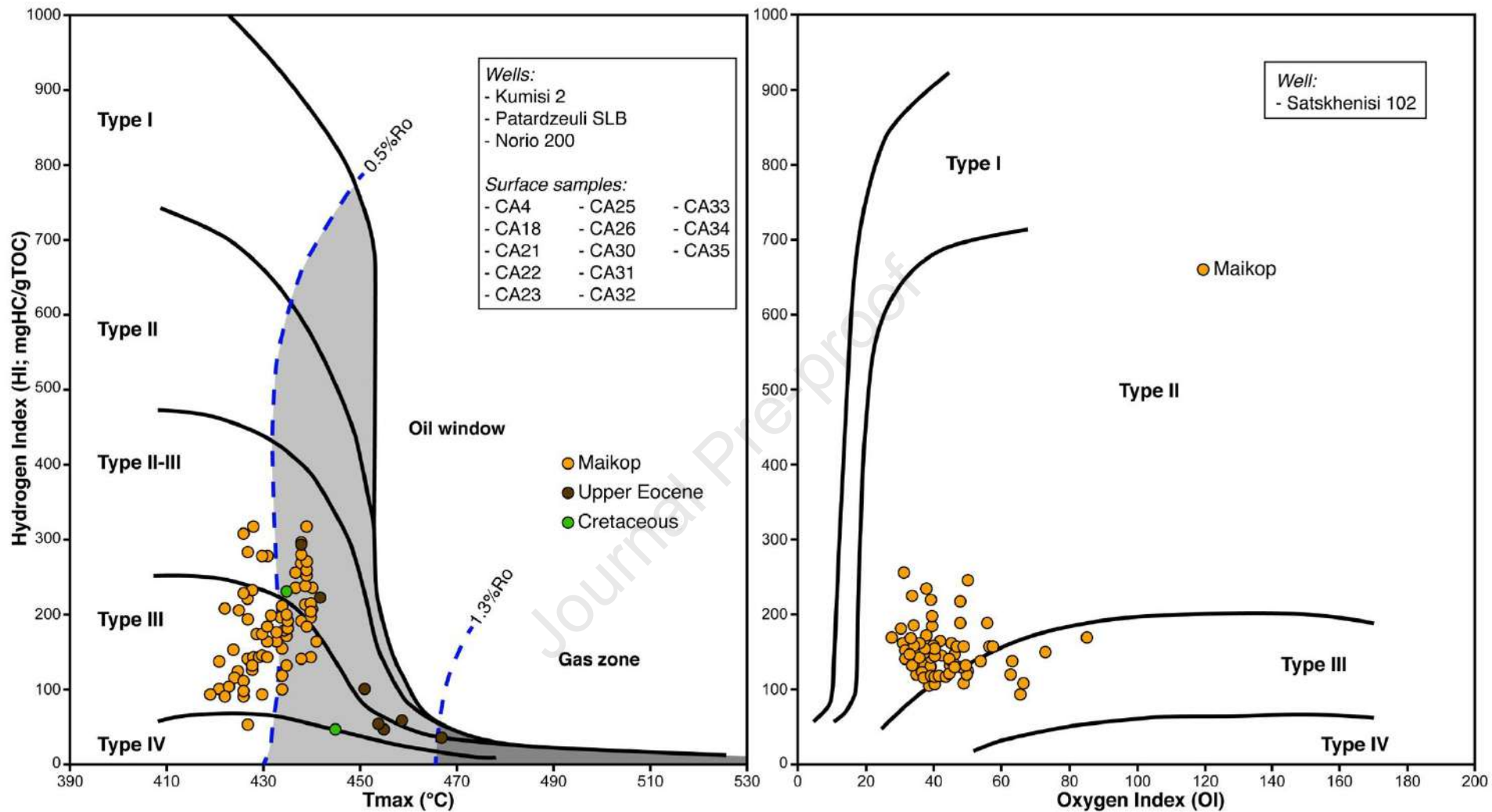


Figure 6: Tmax vs HI diagram (left) for all new data derived from wells and surface samples, subdivided according to their age, presented in this paper, except for well Satskhenisi 102 (Middle Maikop) for which a OI vs HI diagram is presented (right).

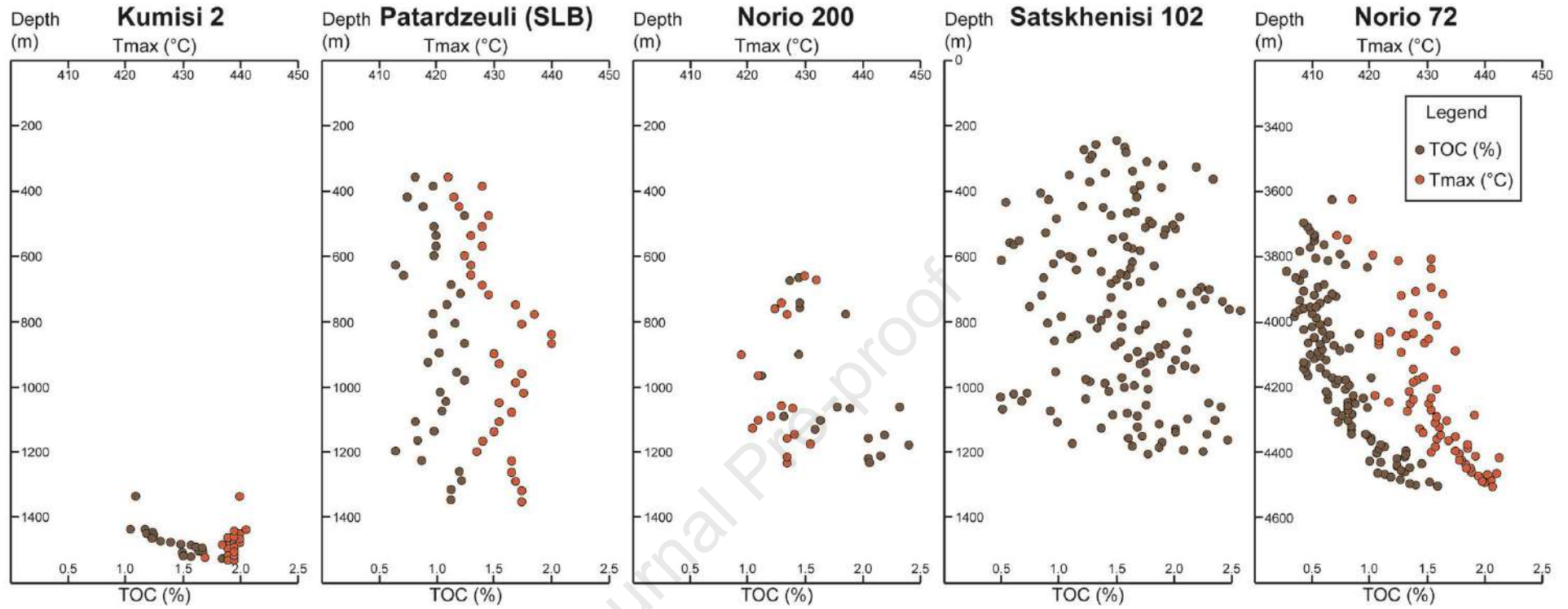


Figure 7: Depth vs Tmax (orange dots) and depth vs TOC (brown dots) plots (upper and lower x axis, respectively) for the five wells having at least two hundred metres of succession. For Satskhenisi 102 well Tmax data are not available.

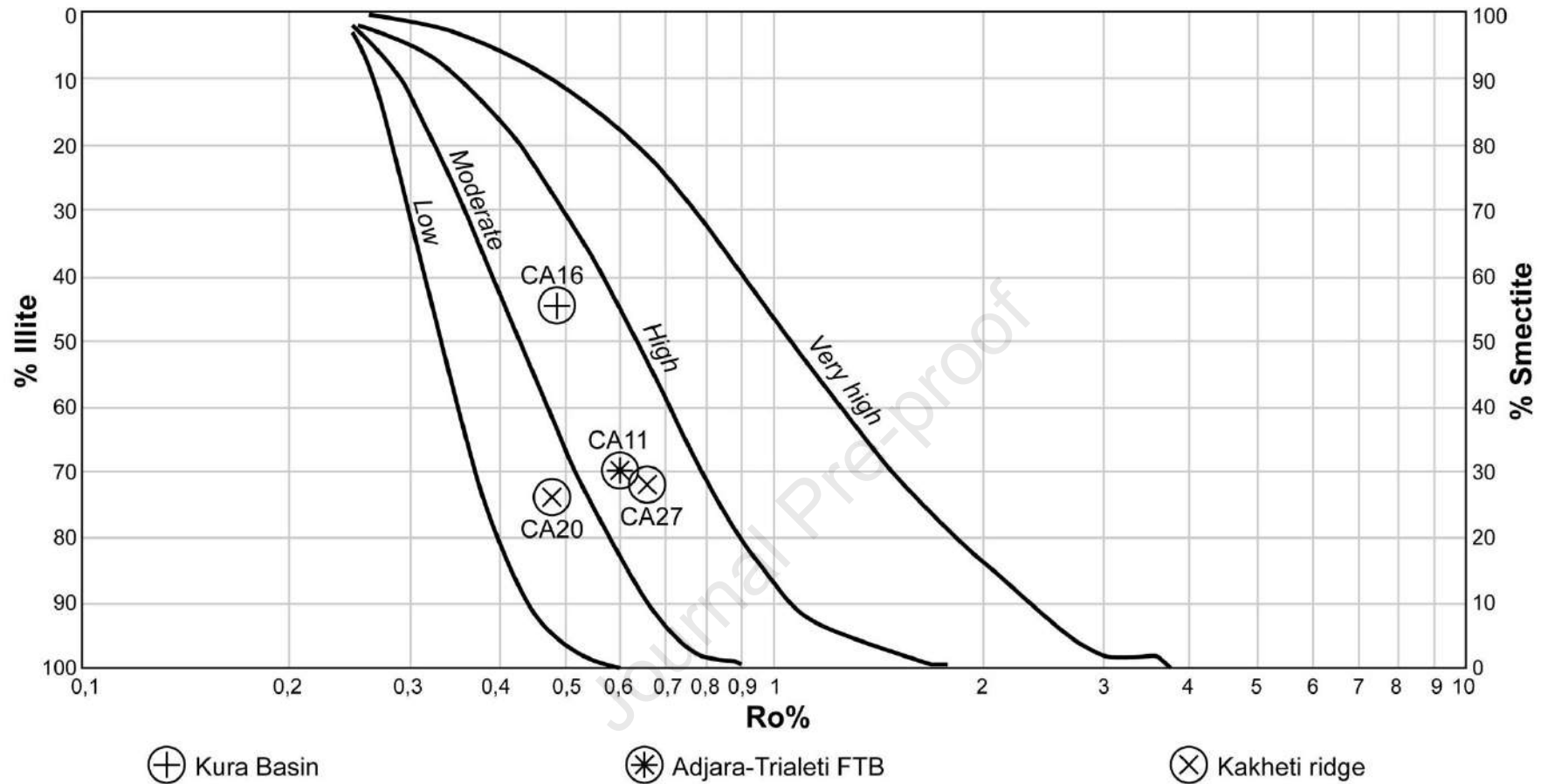


Figure 8: Correlation of VRo% (x axis) and illite% (left y axis) or smectite% (right y axis) in illite-smectite mixed layers, to derive approximate heating rates. Curves indicating heating rates are redrawn and slightly modified after Hillier et al. (1995).

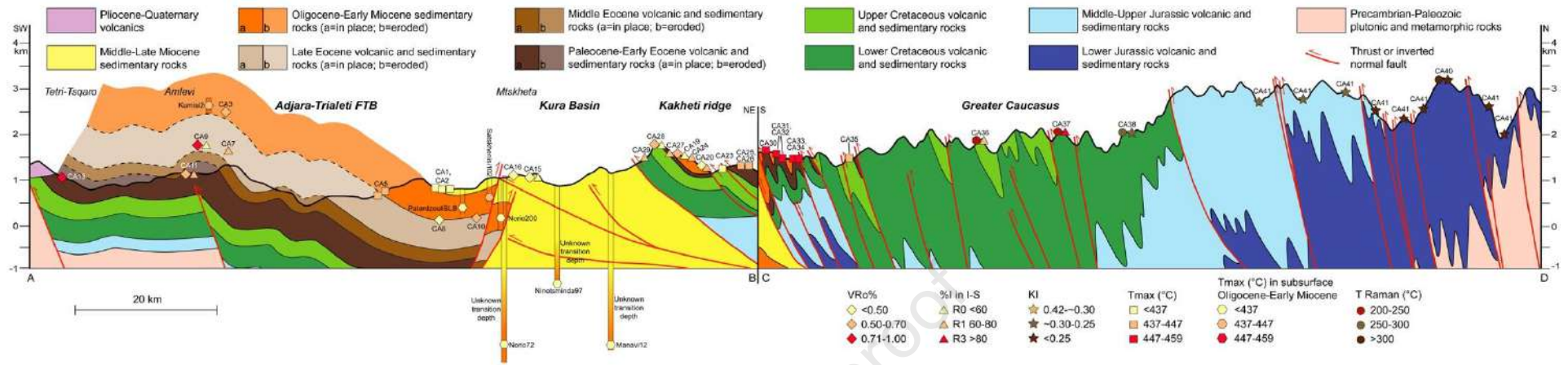


Figure 9: Composite geological section across the study area with original and published synthetic paleo-thermal maturity datasets. Wells and samples are projected on the cross-section line. Redrawn and modified after Alania et al., 2017, 2018; Gusmeo et al., 2021; Mauvilly et al., 2016.



Figure 10: a) Field photograph from the northern side of the Kakhети ridge showing Pliocene-Quaternary flat-lying strata (see dotted black line) overlying Upper Cretaceous tilted rocks (yellow lines); b) Field photograph from the Kura Basin (near Rustavi town) showing tilted Oligocene (Maikop) sandstones and siltstones unconformably overlain (yellow dotted line) by flat-lying Late Pliocene-Early Pleistocene conglomerates. Both images demonstrate that the main phase of deformation within the Kura Basin/Kakhети ridge ended before the Late Pliocene, but uplift continued without tilting. Locations are indicated in Figure 1.

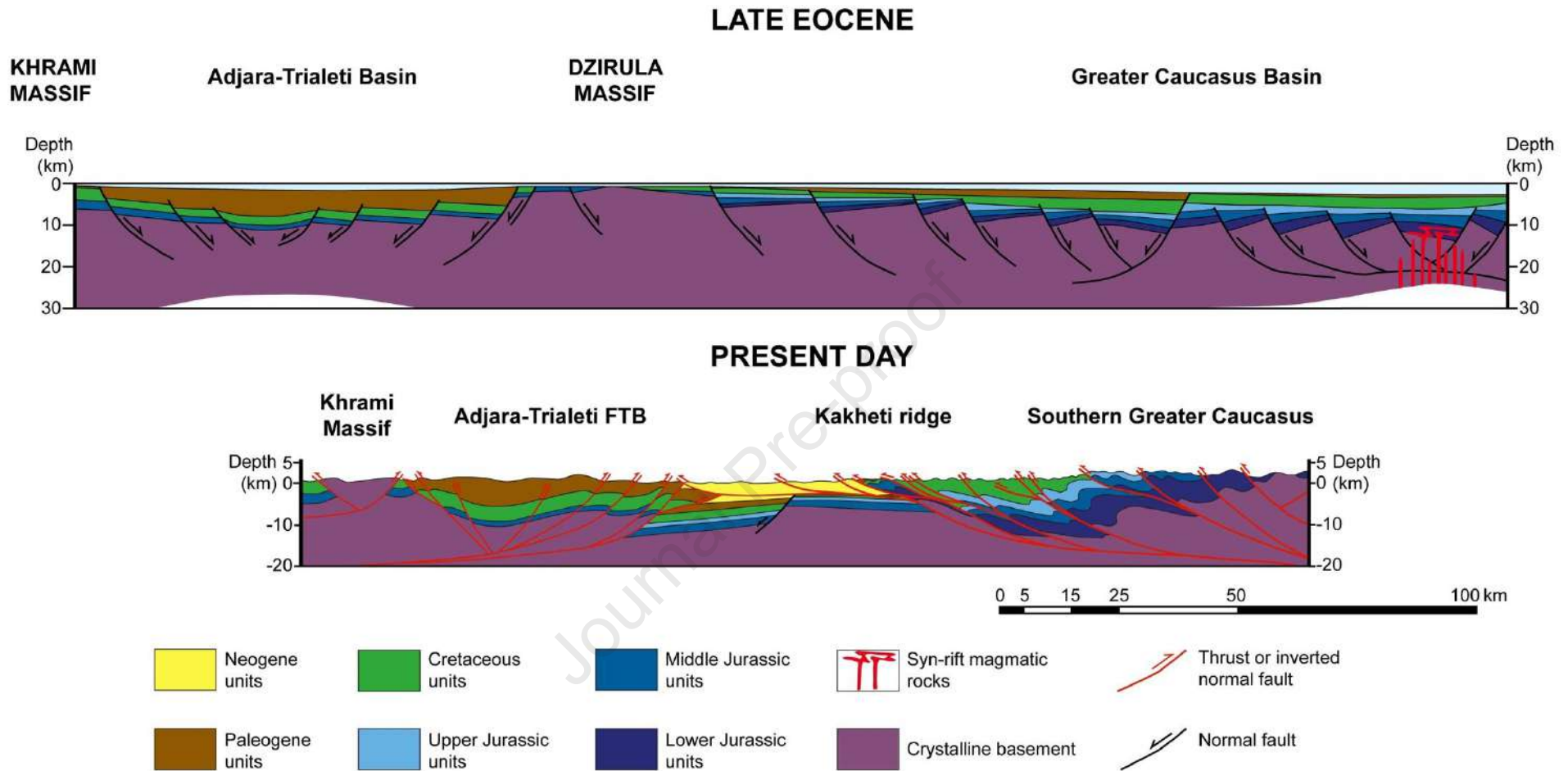


Figure 11: Schematic cartoon of the structural-stratigraphic setting of the study area across the Dzirula Massif, extending to the south in the Adjara-Trialeti FTB, derived from inversion of a Cenozoic back-arc basin, and to the north across the Greater Caucasus, derived from inversion of a Mesozoic rift basin. Redrawn and modified after Alania et al., 2017; Gusmeo et al., 2021; Mauvilly et al., 2016.

Table 1
Thermal maturity data derived from surface samples.

| Sample name | Geological domain | Coordinates | Age | Ro% ± sd (measured fragments) | Tmax (with TOC >0.5) (°C) | Hydrogen Index (mgHC/gTOC) | T Raman (°C) ± sd (n° measurements) | XRD <2µm composition | I% in I-S (R Nr) | Kubler Index | Reference |
|-------------|---------------------|---------------------|-------------------------------|-------------------------------|---------------------------|----------------------------|-------------------------------------|--|------------------|--------------|------------------------|
| CA1 | Adjara-Trialeti FTB | - | Miocene | - | <437 | - | - | - | - | - | Pupp et al., 2018 |
| CA2 | Adjara-Trialeti FTB | - | Late Oligocene-Miocene | - | 429-435 | - | - | - | - | - | Pupp et al., 2018 |
| CA3 | Adjara-Trialeti FTB | 38N 0452009 4616299 | Late Oligocene | 0.51 ± 0.04 (36) | - | - | - | - | - | - | This study |
| CA4 | Adjara-Trialeti FTB | 38N 0325498 4610679 | Oligocene | - | 430 | 93 | - | - | - | - | This study |
| CA5 | Adjara-Trialeti FTB | - | Late Eocene-Oligocene | - | 446-448 | - | - | - | - | - | Pupp et al., 2018 |
| CA6 | Adjara-Trialeti FTB | 38N 0500906 4615010 | Late Eocene | 0.47 ± 0.03 (33) | - | - | - | - | - | - | This study |
| CA7 | Adjara-Trialeti FTB | 38N 0461429 4612640 | Late Eocene | - | - | - | - | I ₃₃ I-S ₄₀ C-S ₂₈ | 76 (R1) | - | This study |
| CA8 | Adjara-Trialeti FTB | 38N 0343226 4618483 | Late Eocene | - | - | - | - | I ₄₅ I-S ₂₄ Ch ₃₁ | 32 (R0) | - | This study |
| CA9 | Adjara-Trialeti FTB | 38N 0452684 4611098 | early Late Eocene | 0.74 ± 0.04 (48) | - | - | - | I ₅ I-S ₂₅ C-S ₃₉ Ch ₃₀ | 27 (R0) | - | This study |
| CA10 | Adjara-Trialeti FTB | 38N 0476497 4631400 | Middle Eocene | 0.50 ± 0.05 (19) | - | - | - | I ₆ C-S ₆₇ Ch ₂₈ | - | - | This study |
| CA11 | Adjara-Trialeti FTB | 38N 0472357 4609370 | Early Eocene | 0.60 ± 0.05 (21) | - | - | - | I ₃₀ I-S ₂₄ C-S ₁₆ Ch ₃₀ | 70 (R1) | - | This study |
| CA12 | Adjara-Trialeti FTB | 38N 0363775 4631968 | Late Paleocene-Early Eocene | 0.60 ± 0.04 (23) | - | - | - | I ₇₂ C-S ₂₂ Ch ₆ | - | - | This study |
| CA13 | Adjara-Trialeti FTB | 38N 0456357 4599540 | latest Paleocene-Early Eocene | 0.77 ± 0.06 (41) | - | - | - | I ₁₇ C-S ₅₀ Ch ₃₃ | - | - | This study |
| CA14 | Kura Basin | 38N 0500909 4595159 | Late Miocene | - | - | - | - | Sm ₈₅ I ₁₀ Ch ₅ | - | - | This study |
| CA15 | Kura Basin | 38N 0512307 4625099 | Middle-Late Miocene | 0.43 ± 0.05 (92) | - | - | - | - | - | - | This study |
| CA16 | Kura Basin | 38N 0512536 4624356 | late Early Miocene | 0.49 ± 0.05 (23) | - | - | - | I ₄₇ I-S ₁₉ K ₃₂ Ch ₂ | 45 (R0) | - | This study |
| CA17 | Kura Basin | 38N 0537395 4591009 | Oligocene-Early Miocene | 0.40 ± 0.03 (31) | - | - | - | - | - | - | This study |
| CA18 | Dzirula Massif/Kura | 38N 0379736 4652510 | Cretaceous | - | <437 | 230 | - | - | - | - | This study |
| CA19 | Kakheti Ridge | 38N 0520846 4635330 | Oligocene-Early Miocene | - | - | - | - | I ₅₄ I-S ₂₆ K ₁₀ Ch ₁₀ | 15 (R0) | - | This study |
| CA20 | Kakheti Ridge | 38N 0529855 4636895 | Oligocene-Early Miocene | 0.48 ± 0.03 (46) | - | - | - | I ₆₄ I-S ₂₄ K ₁₀ Ch ₂ | 74 (R1) | - | This study |
| CA21 | Kakheti Ridge | 38N 0563536 4615033 | Oligocene-Early Miocene | - | 427 | 51 | - | - | - | - | This study |
| CA22 | Kakheti Ridge | 38N 0563937 4615428 | Oligocene-Early Miocene | - | 421 | 99 | - | - | - | - | This study |
| CA23 | Kakheti Ridge | 38N 0495174 4652984 | Late Eocene-Oligocene | - | 434 | 99 | - | - | - | - | This study |
| CA24 | Kakheti Ridge | 38N 0529177 4636408 | Late Eocene-Oligocene | - | - | - | - | I ₅₃ I-S ₂₁ K ₁₅ Ch ₁₂ | 72 (R1) | - | This study |
| CA25 | Kakheti Ridge | 38N 0505210 4652984 | Bartonian-Lower Priabonian | - | 438 | 292 | - | - | - | - | This study |
| CA26 | Kakheti Ridge | 38N 0506810 4656464 | Late Eocene | - | 442 | 222 | - | - | - | - | This study |
| CA27 | Kakheti Ridge | 38N 0529177 4636408 | Paleocene | 0.66 ± 0.05 (7) | - | - | - | I ₆₄ I-S ₂₄ K ₁₀ Ch ₂ | 72 (R1) | - | This study |
| CA28 | Kakheti Ridge | 38N 0524721 4634599 | Maastrichtian | 0.63 ± 0.06 (40) | - | - | - | I ₁₄ I-S ₈₆ | 18 (R0) | - | This study |
| CA29 | Kakheti Ridge | 38N 0528762 4634818 | Hauterivian-Albian | - | - | - | - | I ₃₂ I-S ₉ K ₂₈ Ch ₃₁ | 75 (R1) | - | This study |
| CA30 | Greater Caucasus | 38N 0481457 4663658 | Late Eocene | - | 454 | 53 | - | - | - | - | This study |
| CA31 | Greater Caucasus | 38N 0476888 4666996 | Late Eocene | - | 451 | 101 | - | - | - | - | This study |
| CA32 | Greater Caucasus | 38N 0475448 4667103 | Late Eocene | - | 459 | 57 | - | - | - | - | This study |
| CA33 | Greater Caucasus | 38N 0475448 4667135 | Late Eocene | - | 455 | 46 | - | - | - | - | This study |
| CA34 | Greater Caucasus | 38N 0476401 4667399 | Late Eocene | - | 467 | 36 | - | - | - | - | This study |
| CA35 | Greater Caucasus | 38N 0472854 4669914 | Cenomanian | - | 445 | 45 | - | - | - | - | This study |
| CA36 | Greater Caucasus | 38N 0473963 4680238 | Cenomanian | - | - | - | 216 ± 13 (7) | I ₁ I-S ₂₈ Ch ₇₁ | 80 (R3) | - | This study |
| CA37 | Greater Caucasus | 38N 0474870 4688016 | Aptian-Albian | - | - | - | 228 ± 5 (14) | I ₈₇ I-S ₈ Ch ₅ | 88 (R3) | - | This study |
| CA38 | Greater Caucasus | 38N 0459110 4697896 | Berriasian-Hauterivian | - | - | - | 292 ± 7 (17) | I ₈₈ C-S ₁ Ch ₁₁ | 86 (R3) | 0.29 | This study |
| CA39 | Greater Caucasus | 38N 0544064 4656874 | Oxfordian-Tithonian | - | - | - | 235 ± 22 (16) | I ₈₂ C-S ₁₂ Ch ₆ | - | 0.43 | This study |
| CA40 | Greater Caucasus | 38N 0469216 4720975 | Toarcian-Aalenian | - | - | - | 379 ± 9 (15) | I ₄₆ Ch ₅₄ | - | 0.10 | This study |
| CA41 | Greater Caucasus | - | Lower-Middle Jurassic | - | - | - | - | - | - | <0.30 | Bujakaite et al., 2003 |

List of surface original and published data analysed and discussed in the paper with samples name, geological domain, location, age, selected paleo-thermal parameters (VRo%, Tmax with TOC>0.5, HI, TRaman, I% in I-S, KI) and <2 µm XRD composition. Original and published data are indicated from south to north. Pyrolysis data are derived using various editions of IFP Rock-Eval technology (see Behar et al. (2001) and references therein). For <2 µm XRD composition: Ch = Chlorite, C-S = Chlorite-Smectite mixed layers, I = Illite, I-S = Illite-Smectite mixed layers; K = Kaolinite, Sm = Smectite.

Table 2

Thermal maturity data derived from wells.

| Well | UTM Coordinates | Age | Geological domain | Analysed depth (top-bottom, m) | Analysed thickness (m) | Mean Tmax (°C) (Min, Max) | Mean TOC (Min, Max) | Mean HI (Min, Max) | Nr samples | Sample type | Average Ro% | Reference |
|-------------------|---------------------|-------------------------------|---------------------|--------------------------------|------------------------|---------------------------|----------------------|--------------------|------------|-------------|-------------|-------------|
| Kumisi 2 | 38N 0483141 4607313 | Oligocene-Early Miocene | Adjara-Trialeti FTB | 1340-1535 | 195 | 439 (434, 441) | 1.44 (1.05, 1.95) | 222 (117-318) | 21 | cuttings | - | This study |
| Patardzeuli (SLB) | 38N 0515117 4618697 | Oligocene-Early Miocene | Adjara-Trialeti FTB | 360-1350 | 990 | 431 (422, 440) | 1.01 (0.64, 1.25) | 155 (91-235) | 34 | cuttings | - | This study |
| Satskhenisi 102 | 38N 0507754 4625925 | middle Maikop, Late Oligocene | Adjara-Trialeti FTB | 200-1200 | 1000 | (437-447) | (0.50, 2.40) | (91-259) | 148 | cuttings | 0.3-0.6 | This study |
| Norio 200 | 38N 0494735 4629461 | Oligocene | Kura Basin | 665-1233 | 568 | 426 (419-432) | 1.77 (1.13, 2.41) | 223 (90-318) | 18 | cuttings | <0.5 | This study |
| Norio 72 | 38N 0500117 4630132 | Oligocene | Kura Basin | 3625-4510 | 885 | 429 (415-442) | 0.80 (0.30, 1.61) | 128 (77-208) | 64 | cuttings | 0.5-0.7 | Samsu, 2014 |
| Ninotsminda 97 | 38N 0524029 4624730 | lower Maikop, Early Oligocene | Kura Basin | 2330-2360 | 30 | 422 (421-424) | 0.76 (0.63, 0.82) | 100 (98-102) | 4 | cuttings | <0.5 | Samsu, 2014 |
| Manavi 12 | 38N 0535573 4623781 | Oligocene-Early Miocene | Kura Basin | 3800-3920 | 120 | 424 (407-431) | 4.20 (3.30, 5.40) | 200 (120-279) | 32 | cuttings | <0.5 | Samsu, 2014 |

List of wells used in this study, with wells name, location, age, thickness and depth of the section considered, Tmax, TOC, HI and VRo% data. Pyrolysis data are derived using various editions of IFP Rock-Eval technology (see Behar et al. (2001) and references therein).

Highlights

- *New multi-proxy thermal maturity dataset from Adjara-Trialeti to Greater Caucasus*
- *Thermal maturity jump from Greater Caucasus to Adjara-Trialeti FTB and Kura Basin*
- *Positive inversion of rift basins into Adjara-Trialeti FTB and Greater Caucasus*
- *Thin-skinned deformation in Kura Basin/Kakheti above pre-shortening structural high*

Declaration of interests

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

The authors declare the following financial interests/personal relationships which may be considered as potential competing interests:

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